

A mechanism of internal decadal atlantic ocean variability in a high-resolution coupled climate model

Article

Accepted Version

Menary, M. B., Hodson, D. L. R., Robson, J. I., Sutton, R. T. and Wood, R. A. (2015) A mechanism of internal decadal atlantic ocean variability in a high-resolution coupled climate model. *Journal of Climate*, 28 (19). pp. 7764-7785. ISSN 1520-0442 doi: <https://doi.org/10.1175/JCLI-D-15-0106.1> Available at <https://centaur.reading.ac.uk/40809/>

It is advisable to refer to the publisher's version if you intend to cite from the work. See [Guidance on citing](#).

To link to this article DOI: <http://dx.doi.org/10.1175/JCLI-D-15-0106.1>

Publisher: American Meteorological Society

All outputs in CentAUR are protected by Intellectual Property Rights law, including copyright law. Copyright and IPR is retained by the creators or other copyright holders. Terms and conditions for use of this material are defined in the [End User Agreement](#).

www.reading.ac.uk/centaur

CentAUR

Central Archive at the University of Reading

Reading's research outputs online

A mechanism of internal decadal variability in a high resolution coupled climate model

Matthew B. Menary*

Met Office Hadley Centre, United Kingdom

Daniel L. R. Hodson, Jon I. Robson and Rowan T. Sutton

NCAS-Climate, University of Reading, United Kingdom

Richard A. Wood

Met Office Hadley Centre, United Kingdom

*Corresponding author address: Matthew B. Menary, Met Office Hadley Centre, Met Office,

Fitzroy Road, Exeter, Devon, EX1 3PB. United Kingdom.

E-mail: matthew.menary@metoffice.gov.uk

ABSTRACT

12 The North Atlantic Ocean subpolar gyre (NA SPG) is an important re-
13 gion for initialising decadal climate forecasts. Climate model simulations
14 and palaeo climate reconstructions have indicated that this region could also
15 exhibit large, internally generated variability on decadal timescales. Under-
16 standing these modes of variability, their consistency across models, and the
17 conditions in which they exist, is clearly important for improving the skill of
18 decadal predictions — particularly when these predictions are made with the
19 same underlying climate models. Here we describe and analyse a mode of
20 internal variability in the NA SPG in a state-of-the-art, high resolution, cou-
21 pled climate model. This mode has a period of 17 years and explains 15–30%
22 of the annual variance in related ocean indices. It arises due to the advection
23 of heat content anomalies around the NA SPG. Anomalous circulation drives
24 the variability in the southern half of the NA SPG, whilst mean circulation
25 and anomalous temperatures are important in the northern half. A negative
26 feedback between Labrador Sea temperatures/densities and those in the North
27 Atlantic Current is identified, which allows for the phase reversal. The atmo-
28 sphere is found to act as a positive feedback on to this mode via the North
29 Atlantic Oscillation which itself exhibits a spectral peak at 17 years. Decadal
30 ocean density changes associated with this mode are driven by variations in
31 temperature, rather than salinity — a point which models often disagree on
32 and which we suggest may affect the veracity of the underlying assumptions
33 of anomaly-assimilating decadal prediction methodologies.

34 **1. Introduction**

35 The North Atlantic Ocean has been shown to be a key region for the initialisation of decadal
36 forecasts (Dunstone et al. 2011) and sea surface temperatures (SSTs) in this region are likely im-
37 portant for the climates of the nearby continents of North America and Europe (Rodwell et al.
38 1999). SSTs in the North Atlantic show large multidecadal variability (Knight et al. 2005) which
39 has been linked to drought in the Sahel region (Folland et al. 1986; Zhang and Delworth 2006), At-
40 lantic hurricane formation (Goldenberg et al. 2001; Smith et al. 2010), precipitation over northern
41 Europe (Sutton and Hodson 2005), and the growth and persistence of Arctic sea ice, which could
42 also affect the climate of northern Europe (Screen 2013). In addition to the response to globally
43 increasing greenhouse gas emissions, understanding the natural variability of this region is im-
44 portant in helping to improve the veracity of decadal predictions, which rely in particular on the
45 North Atlantic subpolar gyre (NA SPG) (Dunstone et al. 2011). Indeed, the NA SPG could be an
46 important region in regulating decadal (Hakkinen and Rhines 2004) and longer timescale climate
47 cycles (Kleppin et al. 2015). Due to the paucity of long observational records within the NA SPG
48 much of the mechanistic understanding must be gained through analysis of climate models, some
49 of which is now summarised.

50 There have been many studies investigating the interannual/decadal variability of the NA SPG,
51 which may be useful in adding value to predictions made up to a decade ahead. Given the num-
52 ber of such studies we present here a very brief review. One of the first studies into NA SPG
53 decadal variability proposed a mechanism related to temperature induced gyre changes which ad-
54 vect salinity into the model’s sinking regions and had a periodicity of 50 years in an early coupled
55 climate model (Delworth et al. 1993). Following on from the idealised-ocean work of Frankignoul
56 et al. (1997) this mechanism was ocean-only with the atmosphere providing white noise forcing.

57 However, the agreement between idealised models and fully coupled general circulation models
58 was better in the subtropics than subpolar regions (Frankignoul et al. 1997). Later work found the
59 mechanism of Delworth et al. (1993) in the HadCM3 model (Dong and Sutton 2005) but with a re-
60 duced period of just 25 years with this reduction attributed in part to the removal of flux corrections
61 and the improved representation of surface temperature gradients in the ocean.

62 Using the ECHAM3/LSG model Timmermann et al. (1998) searched for the observationally-
63 based salinity-dominated mechanism of Wohleben and Weaver (1995) related to great salinity
64 anomalies. A periodicity of 35 years was reported and this time the atmosphere was postulated to
65 play a direct, coupled role. However, in parallel work using a coupled model with the same atmo-
66 sphere but a different ocean (ECHAM3/HOPE), Grotzner et al. (1998) also suggested a coupled
67 decadal mode, but with the now even shorter timescale of 17 years, and this time with temperature
68 changes playing an important role. To try and reconcile these differences, Eden and Willebrand
69 (2001) investigated the relative importance of heat and freshwater fluxes related to the North At-
70 lantic Oscillation (NAO) in an ocean-only model and found that, of the two, heat fluxes were more
71 important than freshwater fluxes for the interannual/decadal variability of the NA SPG. However,
72 the coupling between the ocean and atmosphere on multiannual timescales appears to go in both
73 directions (Rodwell et al. 1999; Battisti et al. 1995) suggesting that coupled models are required
74 in order to fully capture the interactions.

75 Disagreements remain over the main contribution to density changes in the NA SPG, along
76 with questions about the degree to which the atmosphere plays a coupled role and the key pro-
77 cesses which set the timescale. However, the general periodicity of simulated multiannual but
78 sub-centennial variability has begun to crystallise (Frankcombe et al. 2010). In addition to the
79 aforementioned works, other studies of this variability within the NA SPG continually find peri-
80 odicities near to 20 years (Visbeck et al. 1998; Watanabe et al. 1999; Holland et al. 2001; Eden

81 and Greatbatch 2003; Dai et al. 2005; Alvarez-Garcia et al. 2008; Danabasoglu 2008; Born and
82 Mignot 2012; Tulloch and Marshall 2012; Sévellec and Fedorov 2013; Escudier et al. 2013; Kwon
83 and Frankignoul 2014). This timescale is sometimes attributed to the basin-crossing timescale of
84 Rossby waves in the NA SPG (Frankcombe et al. 2010) though many studies attribute it to the
85 time to build up sufficient temperature anomalies. Indeed, the role of the forcing from Rossby
86 waves has recently been called into question (MacMartin et al. 2013) and the importance of wave
87 processes in controlling decadal variability is still unclear (Fevrier et al. 2007; Roussenov et al.
88 2008). This approximately 20 year variability is generally separate from centennial variability in
89 the Atlantic, which relies on the long advective timescales to bring anomalies from the tropical
90 Atlantic or Arctic and in which salinity is more consistently the dominant driver of the related
91 density changes (Vellinga and Wu 2004; Jungclauss et al. 2005; Menary et al. 2012). Indeed, the
92 role of salinity in either weakening or strengthening circulations in the NA SPG may depend on
93 timescale (Deshayes et al. 2014).

94 As previously noted, limited instrumental observations within the NA SPG make it hard to detect
95 the existence of decadal variability. However, palaeo reconstructions do suggest increased vari-
96 ance at decadal timescales (Mann et al. 1995) and indeed 20 year variability can be detected on the
97 outskirts of the NA SPG in palaeo proxies (Chylek et al. 2012). Additionally, the relative impor-
98 tance of temperature or salinity variability in real world overturning circulation changes has been
99 investigated. On multiannual timescales, Curry and McCartney (2001) found that the Labrador
100 Sea potential energy anomaly and the overturning were thermally driven — insofar as tempera-
101 tures changed twice as much as salinities in the sinking regions (after scaling by the thermal and
102 haline expansion coefficients).

103 To bring together this previous work with climate models, Figure 1 schematically depicts the
104 studies so far mentioned along with the reported period and whether the proposed mechanism

105 is coupled or ocean-only. Additionally, whether the timescale is reported to be primarily set by
106 either wave processes; the mean circulation strength and the integration of anomalies within the
107 NA SPG; or interaction with the deep western boundary current is noted. Also noted is whether
108 density changes in the Labrador Sea (or model equivalent sinking region) are reported to be tem-
109 perature or salinity dominated. In short, there is much disagreement between the models on the
110 key processes, the details of the mechanism (see above for some examples; the reader is referred
111 to the specific studies for further details), the degree of atmospheric interaction, and the dominant
112 driver of density changes.

113 As analysing decadal variability requires many decades/centuries of integration these previous
114 studies generally use low resolution coupled models ($>1^\circ$ ocean resolution, $>2^\circ$ atmosphere res-
115 olution) or higher resolution ocean-only models. There are reasons to suppose that improved at-
116 mospheric resolution could affect the amplitude of decadal variability (Danabasoglu 2008) whilst
117 improved ocean resolution and associated representation of the Gulf Stream and other bound-
118 ary currents may affect the precise timescales of multiannual/decadal variability (Grotzner et al.
119 1998; Gelderloos et al. 2011; Hodson and Sutton 2012). Higher resolution topography may also
120 be expected to affect the efficacy of wave processes as compared to idealised ocean models with
121 smoothed/no topography (Roussenov et al. 2008; Zhang and Vallis 2007) and improve deep water
122 pathways (Spence et al. 2011). At high ocean resolution eddy induced mixing can be left ex-
123 plicit, rather than parameterised (or the parameterisation significantly turned down), which may
124 impact on the magnitude and variability of ocean heat and freshwater transports (Volkov et al.
125 2008; Treguier et al. 2014). Stronger sea surface temperature gradients, associated with higher
126 ocean resolution, may improve the strength of atmosphere-ocean coupling (Brayshaw et al. 2008).
127 Ultra-high resolution within the Agulhas region has also been shown to affect the variability of the
128 simulated low-latitude Atlantic overturning (Biastoch et al. 2008b).

129 In this study, we document the drivers of NA SPG variability in a new, high-resolution coupled
130 model that represents a rare combination of high resolution (in both ocean and atmosphere) and
131 the multi-century length integration required to analyse decadal timescale modes. We ask: Does
132 high resolution, and the associated processes it allows, affect the nature of simulated decadal
133 variability?

134 The paper is structured as follows: Section 2 describes the model and data used. We then briefly
135 characterise the model in Section 3 before exploring the mechanism of decadal variability in some
136 depth throughout Section 4. The implications of our findings are discussed in Section 5 before
137 conclusions in Section 6.

138 **2. Model description and experimental setup**

139 We examine a prototype of the Met Office Hadley Centre’s state-of-the-art coupled ocean-
140 atmosphere-land ice global environment model, HadGEM3, hereafter referred to as HG3. 460
141 years of near present-day control simulation have been run at high resolution. The atmosphere
142 component is the Met Office Unified Model version 7.7 (Walters et al. 2011). It has a horizontal
143 resolution of N216 (92km at the equator) and 85 levels in the vertical with a model top at 85km
144 with at least 30 levels in or above the stratosphere. The ocean is resolved on the NEMO tripolar
145 grid (0.25°, 75 depth levels, version 3.2, Madec and Coauthors (2008)), with a pole under
146 Antarctica and poles either side of the Arctic Ocean in Asia and North America. The ocean in
147 HG3 was initialised from rest at December 1st using the 2004–2008 time mean EN3 (Ingleby and
148 Huddleston 2007) December-time climatology and subsequently allowed to freely evolve with re-
149 peating 2000 external forcings in the atmosphere. The year 2000 was chosen as it combined a
150 well sampled and recorded set of external forcings with relatively neutral conditions in major cli-

151 mate indices, such as El Niño; for further details of the model configuration and other simulations
152 see Walters et al. (2011).

153 We use observed data from the EN4 objective analysis (Good et al. 2013) which provides infilled,
154 optimally interpolated fields of temperature and salinity on a $1 \times 1^\circ$ grid from 1900 to present-day.
155 EN4 is an updated version of EN3, with improved quality control and error estimates, but was not
156 available when the climate model was initialised. We use the period 1900–2013 to construct a sim-
157 ple climatology for comparison with HG3 and note that the biases in HG3 are large enough (see
158 Section 3a) that the method used to construct the climatology is unlikely to be of first order impor-
159 tance *i.e.* the results are not sensitive to choosing 1900–2013 or 1960–2013 climatologies. Unlike
160 the HG3 model, which is run with interannually constant forcings appropriate for the year 2000,
161 this observational data also includes the effects of all other natural and anthropogenic forcings.

162 HG3 is a precursor to the model used in the Met Office global seasonal forecast,
163 GloSea5 (MacLachlan et al. 2014). GloSea5 will also be similar to the new decadal prediction
164 model. However, there are some differences between the HG3 and GloSea5 models, as GloSea5
165 underwent additional development whilst the HG3 control was running. Most importantly for the
166 present study of the NA SPG is the more diffuse thermocline in the HG3 ocean (NEMO version
167 3.2) as compared to GloSea5 (NEMO version 3.4, see discussion section) (Megann et al. 2014).
168 Despite this the NA SPG biases in upper ocean temperature and salinity (compared to EN4), are
169 small compared to many other coupled climate models used to study NA SPG variability (e.g. Es-
170 cudier et al. (2013), Wang et al. (2014) (for SSTs), see Section 3a). Further details of global
171 mean-state biases within the atmosphere and ocean in HG3 can be found in Walters et al. (2011).

3. Characterising the model

We now examine the NA mean state biases and signal of decadal variability in HG3 in some more detail as a precursor to investigating the mechanisms of variability which exist on top of these biases. In all cases, ‘decadal’ variability refers to 5-year smoothed data, unless otherwise stated.

a. NA SPG Mean state

Mean state biases in top 500m depth averaged temperatures (T500), salinities (S500), and densities (ρ 500) in the NA SPG are less than $\pm 3^{\circ}\text{C}$, $\pm 0.4\text{PSU}$, and $\pm 0.1\text{kg/m}^3$ in the interior NA SPG, with larger $+4^{\circ}\text{C}$, $+0.6\text{PSU}$, and $\pm 0.2\text{kg/m}^3$ biases in the boundary current regions (Figure 2). The temperature and salinity biases are close to being density compensating in the NA SPG but in the subtropical gyre (not the focus of this study) temperature biases dominate resulting in lighter waters. The anomalously cold region in the western SPG, often attributed to the simulated Gulf Stream being too zonal (Kwon et al. 2010), is not as large as in many coupled climate models (Scaife et al. 2011). Warm anomalies exist all along the NA SPG northern boundary currents. These anomalies are associated with reduced ice distribution around southern Greenland and in the Labrador Sea (not shown). Within the NA SPG, simulated deep convection, as estimated from the annual standard deviation in March mixed layer depths (using the mixed layer estimation method of Kara et al. 2000), is located in the Labrador Sea and Irminger Current.

The Atlantic meridional overturning circulation (AMOC) streamfunction in the model is shown in Figure 3a. The zero streamfunction line sits at a depth of 2–3km with the maximum overturning occurring at a depth of approximately 1km. The deeper overturning cell, representing Antarctic Bottom Water (AABW) and Lower North Atlantic Deep Water (LNADW) has a strength of around 3Sv, whereas the shallower AMOC cell, representing the western boundary current and Upper

195 North Atlantic Deep Water (UNADW) has a mean strength of 17Sv for the last 200 years of the
196 simulation.

197 At 26°N it is possible to directly compare the streamfunction in the model to the RAPID observa-
198 tions. The depth of the RAPID overturning maximum is marked with a cross and is approximately
199 200m deeper than in the simulations, which at these depths represents a single model grid cell
200 in the vertical. The depth of the RAPID zero streamfunction line is around 4km, much deeper
201 than simulated. This is not uncommon in models and may be partly an artefact of computing
202 the simulated overturning using the full 3-dimensional velocities (Roberts et al. 2013), although
203 some models do represent a much deeper upper cell (Yeager and Danabasoglu 2012). Neverthe-
204 less, using a ‘RAPID-style’ calculation, after Roberts et al. (2013) with a depth of no motion at
205 4740m, yields a zero streamfunction depth approximately only 250m deeper than using the full
206 3-dimensional velocities. The structure and variability of the streamfunction shallower than this
207 are essentially unchanged. Finally, the NA SPG barotropic streamfunction and associated time
208 series are also shown (Figure 3, b and d) and broadly compare well to observational estimates and
209 high resolution models (Tréguier et al. 2005).

210 Although the depth (1000m) and strength (17Sv) of the maximum of the upper AMOC cell
211 are consistent with observations, the simulated annual variability in this index is weaker than
212 observed. The simulated annual mean AMOC streamfunction at 26.5°N and 1000m depth has a
213 standard deviation of 1.2Sv (0.9Sv if first detrended), compared to an annual standard deviation
214 of 2.3Sv from the 10 years of RAPID data available (Figure 3c). Additionally, the simulated
215 index begins at a low value and then takes several centuries to spin-up to a more stable state more
216 favourably comparable to the observed mean. Although this represents an improvement in this
217 index of the NA circulation, the spin-up of the overturning circulation also results in an increase

218 in northward heat and salt transport within the Atlantic Ocean, causing the NA SPG to drift away
219 from its initialised state to a warmer and saltier state, seen in Figure 2

220 The simulated AMOC index also shows some evidence of multiannual/decadal variability, par-
221 ticularly at the more northerly latitudes of the NA SPG (not shown) in addition to 26N, as in other
222 models (Zhang 2010). The maximum correlation between the simulated AMOC indices at 26.5°N
223 and 50°N occurs when the 50°N index leads by 1 year (correlation of 0.63 with a correlation of
224 0.12 required for significance at the 95% level), suggesting the lower latitude variability is re-
225 sponding to variability further north in the NA SPG. We now move on to examine the decadal
226 variability of the NA SPG in more detail.

227 *b. Signal of decadal variability*

228 The time-mean T500 simulated in HG3 is shown in Figure 4a along with contours at 6 and 10
229 degrees to mark the general shape of the NA SPG. A comparison with observations (EN4) again
230 shows the general warm bias of the NA SPG, particularly towards the edges of the gyre. A power
231 spectrum for T500 over the whole region reveals a significant peak at a period of 16 to 17 years
232 (Figure 4b). This periodicity exists whether using the entire simulation or alternatively removing
233 the first 200 years (not shown), suggesting it is not merely an adjustment process, and so we use
234 the entire time series to maximise the available data. Additionally, the periodicity is not unique to
235 any of the four individual subregions within the NA SPG (dashed regions in Figure 4a); all show
236 a significant peak at 16 to 17 years, as well as the North Atlantic Current (NAC) and Irminger
237 regions (not shown). Indeed, in HG3 many other large scale ocean indices in the NA SPG also
238 reveal peaks in their power spectra at periods of 16 and 17 years, such as SSTs, depth averaged
239 salinities, the AMOC at 50°N, or the strength of the NA SPG itself (as defined by the barotropic
240 streamfunction, *c.f.* Figure 3).

241 In addition to these ocean indices, the NAO index also shows periodicity at 16 to 17 years in its
242 otherwise much whiter spectrum (Figure 4c). This is suggestive of a link from ocean to atmosphere
243 in the region of the NA in which the ocean can impart some of its long term memory on to the
244 atmosphere. Such a feedback might in general be expected to be weaker than similar atmosphere
245 to ocean processes, and related to the strength of the ocean circulation and SST gradients (Nonaka
246 and Xie 2003), and thus detection of this feedback is perhaps at least in part due to the increased
247 signal to noise ratio resulting from the length of the control simulation (though we note this is
248 still short compared to many previous studies with lower resolution models). The mechanistic
249 drivers behind this 17 year mode in the ocean and atmosphere, and the reasons for the particular
250 timescale, are investigated in the next sections; initially characterising the variability in the NA
251 SPG as a whole before targeted analysis of the processes in different regions.

252 **4. Mechanism of decadal variability in the NA SPG**

253 We now diagnose the mechanism of decadal variability within the NA SPG, beginning with a
254 heat budget for the region before investigating how temperature anomalies propagate around the
255 gyre.

256 *a. Heat budget*

257 To begin to understand the variability of T500 in HG3 a heat budget of the NA SPG is diagnosed
258 (Appendix). There is considerable variability in the net heat flux into the NA SPG, the majority
259 of which appears to be attributable to the advective heat fluxes from the south, which results
260 in decadal timescale heat content changes within the NA SPG. Annual and decadal correlations
261 between the total heat flux and net advective fluxes are 0.75 and 0.69 (for annual and decadal data
262 the 95% significance levels, assuming a two-tailed t-test, are 0.12 and 0.37 respectively), whereas

the same for the total heat flux and net surface fluxes are 0.63 and 0.29 (the regression gradients scale similarly) suggesting that particularly on decadal timescales advective heat fluxes dominate the variability. Once within the NA SPG, how do these heat content anomalies evolve?

b. Lagged regression analysis

In order to investigate the spatial characteristics of the heat content variability, lagged regressions of NA SPG T500 on to SST spatially averaged over the NA SPG were performed (Figure 5). T500 anomalies can be seen propagating around the NA SPG: eastwards along the southern boundary whilst spreading into the interior with a timescale of around 4–6 years (notably slower than implied by the mean circulation speed in this region); westwards along the northern edge but south of the Greenland, Iceland, Norwegian (GIN) Seas; into the central Labrador Sea as opposite sign anomalies form in the Gulf Stream region. A similar evolution of anomalies was also found when regressing T500 on to T500 spatial averages over the eastern SPG, NAC region, or Labrador Sea (not shown). The remaining panels will be discussed in Section 4f. Features such as the Reykjanes Ridge can be seen diverting the flow. Although not shown here, there is little evidence of significant amounts of the signal diverting into the GIN Seas in the far northern part of the SPG. The heat content anomalies reach the Labrador Sea from the eastern SPG within a couple of years but several more years are required for the anomalies to spread into the interior SPG. As the heat content anomalies in the Labrador Sea build up so does a cold anomaly in the Gulf Stream/NAC region. The opposite phase of the cycle now begins.

The underlying essence of the cycle is captured by regressing T500 indices in the northern and southern edges of the NA SPG against each other (Figure 6, the same result is also found if removing the spin-up phase, not shown). This shows the southern edge of the NA SPG leading the northern edge by 4–6 years and subsequently lagging changes in the northern edge by 0–2 years

with opposite sign, yielding a half period of 4–8 years and a full period of 8–16 years (constrained here to be even by the use of annual data). The timescale is further increased by 2 years (putting the 16/17 year spectral peak more towards the centre of this range) if a third location in the eastern SPG is added to the regression model, forcing the signal to go via the eastern SPG (by regressing the southern index with an index of the eastern NA SPG, and then regressing the index of the eastern NA SPG with the northern index, not shown), suggesting that the spread in timescales is perhaps related to the superposition of various advective pathways. This decadal mode is generally confined to the top 500m–1km with the exception of the central Labrador Sea where it extends to around 2km (not shown). Decadal variability in the band 10–30 years, encompassing the spectral peak at 17 years, explains >15% of the interannual variability in T500 within the NA SPG, with this value rising to >30% in the centre of the gyre.

The lagged regression analysis leads to two key questions: Firstly, what is controlling the apparent propagation of the heat content anomalies in both a) the Gulf Stream extension/NAC, and b) the northern boundary currents/Irminger Current? Secondly, what is the negative feedback which forms the opposite sign anomaly in the NAC, resulting in a cyclical mechanism and a spectral peak in NA SPG temperatures?

c. Heat content anomalies in the NAC region

To determine what controls the heat content changes on the southern boundary of the NA SPG, the heat budget of the NAC region is examined in more detail. A region was chosen where simulated zonal currents are much stronger than meridional or vertical currents (See Figure 4a, blue box). This simplifies the later interpretation of the decomposition of advective heat fluxes into circulation and temperature components. As noted in the Appendix, it is not possible to close the heat budget precisely, which becomes more apparent for smaller subregions. Table 1 shows the

time mean advective components and net surface heat fluxes for the NAC top 500m. Note that the choice of reference temperature becomes irrelevant when considering the net transport through all faces combined but not when considering open sections (Schauer and Beszczynska-Möller 2009). The most important advective heat fluxes are from the east and west, associated with the mean volume transport through the region from east to west. These advective heat fluxes are approximately balanced by the surface heat fluxes but the sum of the two is not identical to the actual heat content change implied by the in-situ temperatures. This is due to missing diagnostics (See Appendix) and the use of monthly means when computing vT , rather than at each model time step. However, although the means are slightly different, the variability in both time series is well correlated on all timescales at monthly or longer sampling (Table 1). Thus in the ensuing analysis of the variability we treat the budget as sufficiently closed.

The annual and decadal timescale correlations (regression gradients, $W=Watts$) between the advective heat fluxes and net actual heat content changes in the NAC are 0.82 ($0.92 W_{dOHC}/W_{adv}$) and 0.54 ($0.40 W_{dOHC}/W_{adv}$) respectively, compared to 0.43 ($0.92 W_{dOHC}/W_{surf}$) and 0.20 ($0.20 W_{dOHC}/W_{surf}$) for the correlation between surface heat fluxes and the net heat content change (for annual and decadal data the 95% significance levels, assuming a two-tailed t-test, are 0.12 and 0.37 respectively). Thus much of the annual and decadal variability in the heat content changes in the Gulf Stream is associated with advective heat fluxes but there is a role for surface fluxes to modulate these changes, even on decadal timescales. See Appendix 2 for the full heat transport breakdown. Of the advective heat fluxes, the remaining question is whether these are due to the anomalous circulation or anomalous temperature.

For the NAC region it can be seen that slightly more of the advective heat flux variability arises from that due to anomalous circulation advecting mean temperature anomalies ($v'T_0$, Table 3). Although the magnitudes are similar between $v'T_0$ and v_0T' components, the relationship with

the net ocean heat transport (OHT, *i.e.* vT) is not, with $v'T_0$ having a higher positive correlation with OHT. Correlations between $v'T_0$ and OHT are 0.29, 0.36, and 0.42 on monthly, annual, and decadal timescales respectively, compared to 0.00, -0.16, and -0.23 for v_0T' (Table 2). This holds throughout the western half of the southern edge of the NA SPG (not shown), and is associated with a strong background temperature gradient. Thus $v'T_0$ appears to be the dominant advective heat flux in the NAC region on all timescales.

d. Heat content anomalies in the Irminger Current region

The same breakdown of heat content changes into a particular region was applied to the Irminger Current at the entrance to the Labrador Sea (Figure 4a, red box). Similarly to the NAC region, this was chosen where horizontal circulation was well defined in a particular direction and much larger than all orthogonal circulations. The breakdown of heat fluxes into surface, advective, and advective subcomponents is shown in Table 1. Similarly to the Gulf Stream region, the net surface and net advective heat fluxes approximately balance but do not fully explain the directly calculated heat content change. However, as before, the correlation between the sum of the surface and advective components and the flux implied by the actual heat content change is very good on all timescales and so we again treat the budget as sufficiently closed.

For the individual fluxes, on annual timescales, the correlation (regression gradient) between the advective heat fluxes and net heat content changes is 0.56 ($0.56 W_{dOHC}/W_{adv}$), again marginally greater than the correlation between surface heat fluxes and net heat content changes at 0.47 ($0.52 W_{dOHC}/W_{surf}$). On decadal timescales these drop to 0.21 (0.08) and 0.19 (0.09) for advective and surface fluxes respectively. Despite these low decadal correlations, there is still a very large correlation between their sum and the actual net heat content change (Table 1), suggesting that on these decadal timescales no single component of the heat budget can be considered the controlling

influence. This is also indicated by the strong anti-correlation between advective and surface heat fluxes of -0.87 on decadal timescales (for annual and decadal data the 95% significance levels, assuming a two-tailed t-test, are 0.12 and 0.37 respectively).

In contrast to the Gulf Stream region, for the Irminger Current the most important advective heat flux is that due to the mean circulation advecting anomalous temperature ($v_0 T'$, Table 3). $v_0 T'$ has slightly greater variability on all timescales than $v' T_0$ and also shows larger correlations (and regression gradients) with the actual OHT changes on all timescales. Correlations between OHT and $v_0 T'$ for monthly, annual, and decadal variability are 0.83, 0.34, and 0.29 respectively, whereas correlations between OHT and $v' T_0$ are much smaller (Table 2). In our Irminger Current box the zonal currents are an order of magnitude larger than in all other directions, and so we suggest that it is the zonal mean circulation which is playing an important role in moving heat content anomalies from east to west on the northern edge of the NA SPG. Additionally, the simulated mean temperature of the core of the inflow and outflow waters differs by 0.2K in the Irminger Current region, compared to 1.6K for the NAC region, which may help to explain the smaller standard deviations in advective heat fluxes through the Irminger Current region.

In summary, the heat budget for the NA SPG as a whole has been diagnosed and it has been seen that advective heat fluxes play an important role on decadal timescales, but that the relative contributions of circulation and temperature anomalies to the OHT are region specific. We now investigate the remaining question of what controls the negative feedback between Labrador Sea and NAC temperature anomalies.

e. Negative feedback between Labrador Sea and Gulf Stream T500

The anomalous temperatures in the Labrador Sea, which are related to the increased heat flux into the region, affect deep water formation in this region. As noted in Section 1, an assessment of

379 related studies suggests an approximately even split between temperature and salinity control of
380 the Labrador Sea density changes related to increased deep water formation on decadal timescales.
381 Following Delworth et al. (1993) we decompose the simulated density changes in the Labrador
382 Sea into those due to temperature and those due to salinity (Figure 7a). This analysis suggests
383 that in HG3 simulated density changes in the Labrador Sea are due to temperature induced den-
384 sity changes (annual correlation with actual density: 0.64, correlation required for significance at
385 the 95% level, assuming a two-tailed t-test, is 0.12), rather than salinity induced density changes
386 (annual correlation with actual density: -0.06). A lagged correlation analysis confirms that on
387 both annual and decadal timescales density changes are temperature-controlled (Figure 7b). We
388 hypothesise that simulated dense water formation in the Labrador Sea in HG3 contributes to circu-
389 lation anomalies in the NAC region via the creation of an anomalous north-south dynamic height
390 gradient, and as such acts as a negative feedback on to NA SPG temperatures.

391 To examine this hypothesis we calculate a composite difference in the density in a cross section
392 through the NAC which lags the density upstream in the Labrador Sea (Figure 8a). To the north the
393 connection between surface and deep water is revealed with the signal sinking below the surface as
394 it progresses southwards. The north-south density gradient is associated with a change in the local
395 dynamic height (Figure 8b). Despite the negative density anomaly in the south it can be seen that
396 a large part of the dynamic height anomaly is controlled by the northern, positive density anomaly.

397 As the signal of anomalous density spills out of the Labrador Sea this dynamic height gradient
398 increases and is balanced by anomalous shear in the geostrophic velocities (Figure 8c). The mean
399 geostrophic velocity anomaly between the surface and 500m for the pictured transect is 1.2cm/s,
400 increasing to 1.6cm/s for the top 200m only. Thus, an increase in density in the Labrador Sea,
401 associated with a cooling in this region, is followed by a strengthening of the circulation in the
402 NAC, and thus an increase in northward OHT into the NA SPG (with likely also some additional

403 contribution from $v'T'$ as the anomalous circulation acts on anomalously warm, low density sur-
404 face water, *c.f.* Figure 8a). This acts as a negative feedback on the NA SPG temperatures. We now
405 discuss the atmospheric contribution to these ocean feedbacks.

406 *f. The role of the atmosphere*

407 Although the proposed mechanism of decadal (17 year) variability in HG3 has been described
408 mostly in terms of ocean dynamics there are regions where the atmosphere directly forces, or acts
409 as a positive feedback on, the ocean variability.

410 For example, the negative feedback dipole between Labrador Sea and NAC temperatures is
411 reminiscent of the Ekman response to NAO forcing. To quantify the instantaneous (*i.e.* zero
412 lag) impact of the NAO we attempt to isolate its signal similarly to the analysis of Polo et al.
413 (2014). Specifically, the annual mean 3-dimensional ocean density field was regressed onto the
414 wintertime NAO index (both unfiltered, not shown). The direct impact of the NAO was then
415 removed from the density field by scaling the regression pattern by the NAO index and removing
416 the pattern from the density at each time point. Removing the instantaneous NAO-related signal
417 weakens the density/dynamic height and thus geostrophic current response (Figure 8d) calculated
418 in Section 4e, hence suggesting that some of the proposed ocean negative feedback is forced by
419 the atmosphere and not merely an ocean-only process. On annual timescales the magnitude of
420 the current response is reduced by 45% but on longer, decadal timescales the reduction is less
421 stark (13% reduction, shown in Figure 8d). This analysis assumes that the instantaneous impact
422 of the NAO is annually independent and can be linearly separated. To what extent the NAO and
423 ocean temperatures/densities can be seen as one-way forcing from atmosphere to ocean, and to
424 what extent it is actually a coupled feedback (*i.e.* some of the NAO signal is itself forced by the
425 ocean, implied by the spectral peak in the NAO power spectrum Figure 4c), is discussed below.

426 However, the reduction in anomalous circulation response when removing the NAO suggests that
427 atmospheric forcing/the NAO may act to reinforce this ocean feedback.

428 In the northern NA SPG we have previously shown a role for ocean advection in moving heat
429 content anomalies westwards via the mean circulation (Section 4d). At the same time, surface heat
430 fluxes were also shown to be non-negligible. In Figure 5 the SST, T500 (discussed in Section 4b),
431 SHF, Sea Surface Salinity (SSS), Mean Sea Level Pressure (MSLP), and Sea ice are regressed at
432 various lags against NA SPG mean SSTs. The SHF is directed into the ocean and at lag=0,+2
433 is having a cooling effect in the eastern SPG and a warming effect in the western SPG, *i.e.* it is
434 effectively moving heat content anomalies from east to west. This is likely related to the concomi-
435 tant strongly negative NAO anomaly in the MSLP field at the same lags. The actual magnitude
436 of the SHF contribution to the Irminger Current OHC change is similar to the contribution from
437 advective fluxes (Table 1) but, as noted in Section 4d, both are individually quite poorly correlated
438 with the OHC change on multiannual timescales. This is consistent with a mechanism whereby
439 the ocean integrates up the interannually independent forcing from the atmosphere/NAO resulting
440 in decadal timescale variability in ocean heat content. However, the spectral peak in the NAO
441 index (Figure 4c) also implies some ocean to atmosphere forcing.

442 Additionally, in the eastern SPG, the SSTs are anti-correlated with the NAO index, seen both at
443 the lag=0 regression and with the opposite phase at lag=-6. These SSTs are likely a combination of
444 the direct forcing of both 1) the NAO via SHFs and anomalous Ekman and gyre circulation (Hakki-
445 nen and Rhines 2004; Sarafanov et al. 2008) and 2) the advective heat flux associated with the
446 diagnosed mechanism of decadal variability. As noted previously, the simulated NAO shows a
447 spectral peak at 17 years similarly to ocean indices within the NA SPG. It would appear most
448 likely that this atmospheric memory must come from the ocean but unfortunately long enough
449 atmosphere-only experiments with this model are not available to further test this hypothesis.

450 At lag=0 (and with opposite phase at lag=-6), the anomalous NAO-related SHFs show the same
451 sign change over both the Labrador Sea and Gulf Stream/NAC but over the Gulf Stream/NAC they
452 are of the wrong sign to explain the heat content changes (both at the surface and throughout at
453 least the top 500m of the water column). This is consistent with advective heat fluxes playing a
454 much more dominant role in the heat budget of the NAC region (see Section 4c) than the Irminger
455 Current/Labrador Sea region (Section 4d). However, as noted at the beginning of this section, in
456 the NAC region there is a significant portion of the ocean geostrophic circulation (and associated
457 heat transport) response which is itself related to the NAO (*c.f.* Figures 8c and 8d). In short, it is
458 impossible to completely separate the effects of either the atmosphere or ocean.

459 SSS evolves similarly to SST in the NA SPG although the largest changes are associated with
460 movement of the ice edge in the GIN Seas (Figure 5; first, fourth, and sixth columns). In general
461 in the NA SPG, positive salinity anomalies co-vary with positive temperature anomalies in both
462 space and time, again suggesting a role for advective fluxes. NAO-related surface freshwater fluxes
463 are also proposed to be of only secondary importance due to the fact that simulated SSS anomaly
464 magnitudes are independent of the amplitude of the NAO (not shown).

465 Similarly to other large scale variables within the NA SPG, ice edge changes exhibit decadal
466 variability with a spectral peak at a period of 17 years (not shown). However, unlike in similar
467 work with the IPSL model (Escudier et al. 2013) negative sea ice anomalies do not appear to lead
468 cooling in the NA SPG (Figure 5, sixth column). We suggest that in our simulations ice edge
469 changes are primarily a passive response to the temperature dominated decadal variability within
470 the NA SPG, perhaps again via the NAO (Deser et al. 2000), rather than a direct driver of this
471 variability.

472 *g. Summary of the proposed mechanism*

473 The mechanism of decadal (17 year) variability simulated in the NA SPG T500 and SSTs is
474 summarised in Figure 9. Positive circulation anomalies in the southern part of the SPG move heat
475 eastwards and northwards into the eastern SPG with a timescale of around 5 years (orange). These
476 heat content anomalies are then transported by the mean circulation around the northern edge of
477 the SPG with a timescale of around 2 years (red). In the Labrador Sea these anomalies affect the
478 stability of the water column. These negative density anomalies, associated with reduced deep
479 water formation, spill out from the Labrador Sea into the SPG, deepening as they go. In the region
480 north of the Gulf Stream these negative density anomalies affect the north-south density gradient
481 and induce geostrophic circulation anomalies weakening the NAC. The weaker circulation reduces
482 ocean heat transport and acts to cool the NA SPG (blue). The phase of the oscillation is thus
483 reversed.

484 The postulated role of the atmosphere is also noted (black dashed lines in Figure 9): As tem-
485 perature anomalies build up in the eastern SPG the atmosphere acts to strengthen these anomalies.
486 When the east of the NA SPG is anomalously warm or cold SHFs also act to move the ocean
487 heat content anomaly westwards. Lastly, in the region of the Labrador Sea/Gulf Stream temper-
488 ature (density) dipole the NAO is associated with around 13% of the ocean-circulation feedback
489 (calculated in Section 4e and shown in Figure 8d).

490 We now discuss the implications of our work and similarities between it and previous studies.

491 **5. Discussion**

492 In the context of the brief literature summary in Section 1, and the schematic illustration
493 presented in Figure 1, our simulations broadly fall into a temperature-dominated regime in the
494 Labrador Sea in which the mechanism could be described as ‘Ocean*’ (where the asterisk implies

495 a positive feedback between the NAO and SSTs may be amplifying the mode, after Figure 1).
496 The timescale is set in part by mean circulation speeds in the northern SPG but with a transition
497 to anomalous circulation in the southern SPG — although it is not clear from the simulations
498 precisely where this transition occurs.

499 The simulated timescales between changes in the Labrador Sea, NAC and eastern SPG have been
500 attributed to advective processes. However, confounding this are wave processes which are also
501 weakly detectable within the model. Analysis of the deep density field (1500–3000m, not shown)
502 reveals signals characteristic of boundary waves propagating from the Labrador Sea to the equator;
503 propagating along the equator to the eastern boundary; subsequently propagating north and south
504 along the eastern boundary, all the while radiating Rossby waves westwards. The evolution is
505 very similar to that found in the idealised model of Johnson and Marshall (2002) and yield a lag
506 between the Labrador Sea and eastern SPG of 5 years, broadly similar to that due to advection.
507 Although detectable, these wave signals require heavy filtering of the deep density field whilst the
508 proposed mechanism exists mainly in the top 1km. We can only conclude that wave processes
509 may play an additional role in our simulated variability but the magnitude of this is unclear. We
510 also note that the relatively diffuse thermocline in HG3 (Megann et al. 2014) may act to dampen
511 these wave processes (Grotzner et al. 1998) as compared to the updated seasonal forecast model,
512 GloSea5 (which will be similar to the new Met Office decadal prediction model).

513 Despite the lagged regression analysis used in this study, and its ubiquity within studies of
514 decadal variability within climate models, there are some hints from the present work that the
515 proposed mechanism may be asymmetric. This asymmetry is manifest in the timescales of various
516 phases of the cycle being also dependant on the sign of the anomaly; *i.e.* the same processes are
517 at work in opposite phases of the mechanism but may evolve with different timescales. Some
518 evidence for this can be seen in Figure 5 in which all the fields reverse sign over 6–8 years,

519 implying a periodicity of 12–16 years, and yet the spectral peak occurs at the upper end of this
520 at 16 to 17 years. If we construct lagged composites of the T500 (or SST) field based on the
521 top/bottom 10% of phases of the SST index (not shown) we find a reversal timescale of 10 years
522 following a high SST phase, but a reversal time of 6 years following a low SST phase. This
523 asymmetrical timescale doesn't appear to be directly due to the effect of heat transport by the
524 anomalous circulation in the southern SPG ($v'T_0$, see Section 4c) as the lags between the NAC and
525 eastern SPG are the same timescale in both phases. It is important to note though that constructing
526 composites, which only use 20% of the total data, reduces the number of degrees of freedom.

527 This asymmetry also appears evident in the coupled simulation when compositing MSLP based
528 on high/low phases of the SST index. Although both MSLP patterns, composited against posi-
529 tive SSTs (Figure 10a) and negative SSTs (Figure 10b), show significant MSLP anomalies, the
530 magnitude and precise structure are clearly different, with only the negative SST composite as-
531 sociated with the canonical NAO pattern (Figure 10b). Additionally, atmosphere-only sensitivity
532 experiments (not shown) suggest a stronger coupling in the NA SPG between anomalously pos-
533 itive NAO/negative SSTs than anomalously negative NAO/positive SSTs. We plan to investigate
534 the asymmetry further in a separate study.

535 *Comparison with observations and other models*

536 It is difficult to that the prove the mode of variability reported here is inconsistent with observa-
537 tional data due to the paucity of observational records in the NA SPG, particularly in the northern
538 half, and the presence of confounding additional transient forcings in the observational record.
539 However, palaeo proxies from the NA SPG suggest there is 20 year variability in some indices
540 in the region (Sicre et al. 2008; Chylek et al. 2012), although it must be noted that there is dis-
541 agreement on the spectral characteristics of all proxies (Mann et al. 1995). The specific elements

542 of our proposed mechanism (anomalous circulation OHT in the southern part of the NA SPG,
 543 mean circulation OHT in the northern part, a negative feedback between Labrador Sea and NAC
 544 temperatures) are also broadly consistent with the observational literature. For example, there
 545 are some similarities to the anti-correlated relationship between Labrador Sea and NAC tempera-
 546 tures/transport seen in observations (Curry and McCartney 2001). This observational work also
 547 highlights the significant role of the NAO in this relationship as well as the dominant role for tem-
 548 perature (as opposed to salinity) in driving these changes. We note that as a result of the northern
 549 NA SPG warm bias in HG3 there is less ice in the mean, which may detrimentally affect the ability
 550 of ice/freshwater fluxes to affect the decadal variability. In models where the NA SPG mean state
 551 bias is cold, feedbacks involving ice and freshwater fluxes have been shown to be crucial to the
 552 diagnosed decadal variability (Escudier et al. 2013).

553 Although it is difficult to isolate the precise mechanisms by which increased ocean or atmo-
 554 sphere resolution may have altered our results — without a parallel set of low resolution simula-
 555 tions within the same model framework — there are specific features of the decadal variability that
 556 are likely to be affected by enhanced resolution. For example, our proposed mechanism of NA
 557 SPG decadal variability suggests a prominent role for boundary currents, which may be improved
 558 by higher resolution (Grotzner et al. 1998; Gelderloos et al. 2011). Additionally, the increased
 559 atmospheric resolution (which represents the main computational burden for the coupled model)
 560 may affect the innate atmospheric variability over the North Atlantic (Matsueda et al. 2009), while
 561 the role of the atmosphere may also be modulated by the improved ocean resolution (Scaife et al.
 562 2011). Recent work comparing 1° , 0.25° , and $1/12^\circ$ resolution simulations with the same under-
 563 lying model suggest that 0.25° is a significant improvement over 1° , in terms of the biases in NA
 564 SPG SSTs and the location of the Gulf Stream, but that there are still further improvements to be
 565 had at even higher resolution (Marzocchi et al. 2015).

Similar to our findings, recent ultra-high resolution ($1/12^\circ$ horizontal resolution) eddy resolving ocean-only model studies show that much of the OHT into the eastern NA SPG occurs in the near surface (but below the Ekman layer) originating in the subtropics (25% of virtual floats at 500m, compared to less than 10% at 50m or 1000m (Burkholder and Lozier 2011, 2014)). In addition, the role of anomalous circulation transporting the mean temperature gradient in the southern part of the NA SPG is indirectly supported by these ocean-only simulations, which find that the mean circulation is unable to explain the slow timescale by which temperature anomalies move from the subtropics to the eastern SPG. Important for decadal variability in our simulations are advective heat fluxes from the southern edge of the NA SPG due to the anomalous circulation ($v'T_0$). However, annual variability (*i.e.* the anomaly) in the AMOC at 26.5°N from 10 years of RAPID data is approximately double the annual standard deviation in HG3. Thus, if the proposed mechanism exists in reality then it could be expected to have a larger amplitude or faster timescale.

The mechanism we have presented has a timescale of 17 years, similar to the 20 years found in the IPSL-CM5A-LR model recently investigated by Escudier et al. (2013). However, a similar timescale does not imply the same mechanism: see for example an identical 17 year timescale but different mechanism reported by Born and Mignot (2012). The present study reports a mode of variability where temperature dominates the density budget, whereas Escudier et al. (2013) report a mode in which freshwater/salinity fluxes have an important role. Indeed, salinity advection within the SPG has been proposed as a cause of bistability in the SPG (Born et al. 2013), albeit on longer timescales. It is intuitive that whether the density budget is dominated by temperature or salinity would affect whether a strengthening northward circulation acted as a positive or negative feedback — but why are NA SPG density changes differently controlled in the two models?

One hypothesis is that the nature of the biases (compared to observations) affects the variability as the non-linear equation of state for density becomes increasingly salinity dominated at cooler

temperatures. To estimate this effect we compute the density change in the Irminger Current region, mechanistically important in both studies, for a one standard deviation change in temperature or salinity (whilst keeping the other of salinity or temperature at climatological values) in both HG3 and the IPSL-CM5A-LR model as well as an observational estimate from EN4 (Table 4). In HG3, such a temperature change has double the impact on density than a change in salinity. This is not the case in the IPSL model where salinity changes are found to be more important (the relative magnitudes are unchanged if we remove the spin-up in HG3, not shown). The EN4 data suggest that the real world may be in a temperature dominated regime, similar to HG3. This points to there being some relationship between the NA SPG mean state biases of a given model and the subsequently diagnosed mechanisms of decadal variability. Note that this cursory analysis merely compares mean states and variability, and does not explicitly investigate whether density variability is temperature- or salinity-controlled. Nevertheless, one implication of this would be that decadal prediction studies using anomaly-assimilation methods, in which the mean state biases are implicitly assumed to be independent of the variability, would need to re-evaluate the validity of this assumption (Robson 2010). We plan to investigate this relationship in more detail in a forthcoming study (Menary et al. 2015).

6. Conclusions

We have analysed a decadal mode of variability in the near surface (top 500m) of the North Atlantic subpolar gyre (NA SPG) in a 460 year control simulation with a version of the high resolution coupled climate model HadGEM3 (HG3).

- The mode of variability involves the propagation of heat content anomalies around the NA SPG with a periodicity of around 17 years.

- Simulated decadal variability (between 10 to 30 years) in the NA SPG explains more than 15% of the annual mean variance in top 500m depth averaged temperatures. This rises to >30% of the variance within the interior NA SPG and Labrador Sea.
- The simulated NA SPG heat budget is dominated by advective, rather than surface, heat fluxes on decadal timescales, with advection from the subtropics playing the primary role. For the specific regions of interest, namely the Irminger Current and North Atlantic Current (NAC), advective fluxes were also found to dominate. The large depth extent of the mode is also consistent with an important role for advection (Saravanan and McWilliams 1998).
- The role of mean or anomalous circulation in transporting heat content anomalies was found to vary with region: Anomalous circulation dominated the variability in the NAC with mean circulation, and hence temperature anomalies, dominant in the Irminger Current region.
- A negative feedback, required for the mechanism to result in a spectral peak, occurs between the Labrador Sea and NAC. Here, density anomalies spill out of the Labrador Sea resulting in a dynamic height gradient across the NAC/Labrador Sea which induces vertical shear in the geostrophic currents. These current anomalies result in heat transport anomalies which reverse the cycle. The density changes are temperature, rather than salinity, driven.
- Variability in the NAO directly contributes to various stages of the mechanism as well as showing signs of responding to ocean variability. Removing the North Atlantic Oscillation (NAO) signal from the negative feedback between Labrador Sea and NAC temperatures/densities (see Section 4f) shows about 45% of the geostrophic current speed feedback is related to the NAO on annual timescales but that on decadal timescales the ocean feedback still dominates. The atmosphere also acts to reinforce temperature anomalies in the eastern

NA SPG and aid their westward propagation in the northern SPG. The proposed mechanism is summarised in Figure 9.

- Whether density changes are temperature or salinity controlled effects where, and how, negative feedbacks can occur. This may also be expected to affect the particular mechanism simulated in the model. This could have important implications for decadal prediction studies that use the method of anomaly-assimilation and prediction, in which the future evolution of the model is assumed to be independent of the mean state — an assumption which we suggest may not be valid.

A modified version of the model presented here will be used as part of the Met Office decadal prediction system and analyses such as we have presented will be important in developing and evaluating such systems. Given the relationship between resolution and the improved realisation of particular processes, as well as mean state biases, further high resolution coupled model studies would be valuable in testing whether these results are robust.

Acknowledgments. MM and RW were supported by the Joint DECC and Defra Hadley Centre Climate Programme, DECC/Defra (GA01101). DH and RS were supported by NERC through the National Centre for Atmospheric Science (NCAS). JR was supported by the Projet Prévisibilité Climatique Decennale (PRECLIDE) and the Seasonal-to-decadal climate Prediction for the improvement of European Climate Service project (SPECS) GA 308378. Data from the RAPID-WATCH MOC monitoring project are funded by the Natural Environment Research Council and are freely available from www.rapid.ac.uk/rapidmoc. The authors would like to thank Matthew Mizielinski for initially setting up the simulations. For access to, and time on, the MONSooN supercomputer we acknowledge support from the Met Office and the Natural Environment Research Council (NERC).

APPENDIX

Heat budget

The basin-wide, full depth NA SPG heat budget is shown in Figure 11 for the latitude range 53–73°N. Due to the lack of availability of the correct ocean diagnostics at high enough output frequency (precluded by the expense of storing high resolution atmosphere and ocean data), the heat budget of the NA SPG does not close perfectly (*c.f.* red and black lines in Figure 11a). However, the error is negligible, less than 1% of the net surface fluxes of the region. Sensitivity tests where all output diagnostics were computed online and stored revealed that horizontal diffusion was the most important missing heat flux. The heat budget of the NEMO ocean is further complicated by the use of a linear free surface with variable volume which sits on top of the fixed volume ocean grid cells and a heat flux between the two. For further details of the precise formulation of the heat budget within the NEMO ocean model see Madec and Coauthors (2008).

The heat budget (ocean heat content (*OHC*) rate of change) of the NA SPG can first be broken down into advective (Q_{adv}) and surface fluxes (Q_{surf}), which add together to give the net heat flux *into* the volume. There are also additional smaller heat fluxes from the ice to ocean, between the linear free surface and fixed ocean volume, and from geothermal heating of the abyssal ocean, particularly in the vicinity of the mid Atlantic ridge, Q_{ice} , Q_{free} , and Q_{geo} respectively:

$$\frac{dOHC}{dt} = Q_{adv} + Q_{surf} + Q_{ice} + Q_{free} + Q_{geo} \quad (A1)$$

The advective fluxes can be further broken down into fluxes from the north (*OHTN*) and south (*OHTS*, positive northward) whilst the surface fluxes can be broken down into the shortwave (solar), longwave, latent, and sensible heat fluxes:

$$Q_{adv} = OHTS - OHTN \quad (A2)$$

$$Q_{surf} = Q_{SW} + Q_{LW} + Q_{lat} + Q_{sens} \quad (A3)$$

677 This reveals that *OHTS* dominates the variability in advective heat fluxes: Using annual data,
 678 the standard deviation of *OHTS* is 28PW, compared to 17PW for *OHTN*. The variability in *OHTS*
 679 is split between vertical ‘AMOC’ and horizontal ‘gyre’ heat transport variability at these latitudes
 680 (annual correlation between *OHTS* and *OHT_{AMOC}* is 0.74, and between *OHTS* and *OHT_{gyre}* is
 681 0.88). The surface fluxes (directed into the ocean) are dominated by shortwave (solar) heating of
 682 the NA SPG, whereas longwave, latent, and sensible heat fluxes represent net heat loss from the
 683 NA SPG.

684 To investigate the relative magnitudes of their variability, and any trends, the mean of each heat
 685 flux over the years 22–42 is removed (Figure 11b). Rather than remove the full time mean, remov-
 686 ing the mean from just the period soon after the model was initialised serves to additionally show
 687 how the heat fluxes diverge. Net advective heat fluxes into the region are increasing throughout the
 688 period, balanced largely by increasing surface heat flux loss, but with some residual heating of the
 689 NA SPG. The advective heat flux trend is dominated by the increase in heat flux from the south,
 690 which is due to the strengthening AMOC (Figure 3c), with much of this heat lost via latent heat
 691 loss as well as longwave emission. The rate of net warming is highest in the first century, which
 692 is also why the net heat flux appears to be below zero for the remainder of the time, *i.e.* the net
 693 warming rate is slower in the subsequent years.

694 APPENDIX 2

695 Heat budget breakdown

Previously Dong and Sutton (2001), showed the advective heat budget for a region in approximate long term equilibrium could be estimated by considering perturbations around a long term mean as:

$$q(t) = \bar{q} + q'(t) \quad (21)$$

where $q(t)$ is some quantity varying in time, \bar{q} is its time mean, and $q'(t)$ is the anomaly in q at each time, t . For the case of the net advective heat transport convergence, replacing q with both v (velocity) and T (temperature) and dropping the (t) on the right hand side for clarity gives:

$$OHT(t) = \rho \times c_p \times \int \left(\bar{v}\bar{T} + v'T' + \bar{v}T' + v'\bar{T} \right) dA \quad (22)$$

where ρ is density, c_p is the heat capacity of sea water, $\bar{v}\bar{T}$ is a constant (the mean heat transport when multiplied by ρc_p), $v'T'$ is the heat transport due to co-variances in circulation and temperature (and is usually but not always small for large enough areas), $\bar{v}T'$ is the heat transport by the mean circulation, $v'\bar{T}$ is the heat transport by the anomalous circulation, and dA indicates integrating over all faces enclosing the volume. vT is estimated at horizontal and vertical faces.

However, as previously mentioned, there is a trend in the NA SPG temperatures, and so the breakdown of the heat budget is made more complicated. For the case of a known trend in one or more of these parameters (*e.g.* temperature) the q' term will not just represent the annual/decadal anomaly but will also have a component due to the trend with the relative contributions to q' varying in size depending on the magnitude of the trend compared to the magnitude of the variability. Thus q must be detrended prior to combining the terms together, *e.g.*

$$q(t) = q_0 + q_1 \times t + q'(t) \quad (23)$$

713 where q_0 is the intercept, q_1 is the linear trend multiplied by time, t , and q' is the perturbation
 714 from this trend. Setting $t = 0$ at the mid point of the linearly trending time series results in q_0 also
 715 representing the mean (previously \bar{q}). This results in the OHT becoming an equation of nine terms
 716 (as we detrend v as well due to the trend in the AMOC, Figure 3c):

$$OHT(t) = \rho \times c_p \times \int \left(v_0 T_0 + v_0 T_1 t + v_0 T' + v_1 T_0 \right. \\ \left. + v_1 T_1 t^2 + v_1 t T' + v' T_0 + v' T_1 t + v' T' \right) dA \quad (24)$$

717 where the terms inside the integral on the right hand side respectively refer to: 1) The time mean
 718 OHT, 2) the interaction between the temperature trend and the mean circulation, 3) the OHT by the
 719 mean circulation, 4) the interaction between the mean temperature and the trend in circulation, 5)
 720 the interaction between the trends in both circulation and temperature, 6) the interaction between
 721 the trend in circulation and the anomalous temperatures, 7) the OHT by the anomalous circulation,
 722 8) the interaction between the anomalous circulation and the trend in temperatures, and 9) the OHT
 723 due to co-variances in circulation and temperature. Analysis of these components reveals a non-
 724 zero contribution from trend-related terms to the advective heat budget variability, but this is much
 725 smaller than the mean and anomalous circulation terms ($v_0 T'$ and $v' T_0$) and as so we focus on these
 726 latter circulation terms.

727 References

- 728 Alvarez-Garcia, F., M. Latif, and A. Biastoch, 2008: On multidecadal and quasi-decadal north
 729 atlantic variability. *Journal of Climate*, **21** (14), 3433–3452.
- 730 Battisti, D., U. Bhatt, and M. Alexander, 1995: A modeling study of the interannual variability
 731 in the wintertime North Atlantic Ocean. *Journal Of Climate*, **8** (12), 3067–3083, doi:{10.1175/

1520-0442(1995)008\textless3067:AMSOTI\textgreater2.0.CO;2}.

Biastoch, A., C. W. Boening, J. Getzlaff, J.-M. Molines, and G. Madec, 2008a: Causes of interannual-decadal variability in the meridional overturning circulation of the midlatitude north atlantic ocean. *Journal Of Climate*, **21 (24)**, 6599–6615, doi:{10.1175/2008JCLI2404.1}.

Biastoch, A., C. W. Boening, and J. R. E. Lutjeharms, 2008b: Agulhas leakage dynamics affects decadal variability in Atlantic overturning circulation. *Nature*, **456 (7221)**, 489–492, doi:{10.1038/nature07426}.

Born, A., and J. Mignot, 2012: Dynamics of decadal variability in the atlantic subpolar gyre: a stochastically forced oscillator. *Climate dynamics*, **39 (1-2)**, 461–474.

Born, A., T. F. Stocker, C. C. Raible, and A. Levermann, 2013: Is the atlantic subpolar gyre bistable in comprehensive coupled climate models? *Climate dynamics*, **40 (11-12)**, 2993–3007.

Brayshaw, D. J., B. Hoskins, and M. Blackburn, 2008: The storm-track response to idealized sst perturbations in an aquaplanet gcm. *Journal of the Atmospheric Sciences*, **65 (9)**, 2842–2860.

Burkholder, K. C., and M. S. Lozier, 2011: Mid-depth lagrangian pathways in the north atlantic and their impact on the salinity of the eastern subpolar gyre. *Deep Sea Research Part I: Oceanographic Research Papers*, **58 (12)**, 1196 – 1204, doi:http://dx.doi.org/10.1016/j.dsr.2011.08.007, URL <http://www.sciencedirect.com/science/article/pii/S09670637110%01531>.

Burkholder, K. C., and M. S. Lozier, 2014: Tracing the pathways of the upper limb of the north atlantic meridional overturning circulation. *Geophysical Research Letters*, **41 (12)**, 4254–4260.

Cabanes, C., T. Lee, and L.-L. Fu, 2008: Mechanisms of interannual variations of the meridional overturning circulation of the North Atlantic Ocean. *Journal Of Physical Oceanography*, **38 (2)**, 467–480, doi:{10.1175/2007JPO3726.1}.

Chylek, P., C. Folland, L. Frankcombe, H. Dijkstra, G. Lesins, and M. Dubey, 2012: Greenland ice core evidence for spatial and temporal variability of the atlantic multidecadal oscillation. *Geophysical Research Letters*, **39** (9).

Curry, R., and M. McCartney, 2001: Ocean gyre circulation changes associated with the North Atlantic Oscillation. *Journal Of Physical Oceanography*, **31** (12), 3374–3400, doi:{10.1175/1520-0485(2001)031\textless3374:OGCCAW\textgreater2.0.CO;2}.

Dai, A., A. Hu, G. Meehl, W. Washington, and W. Strand, 2005: Atlantic thermohaline circulation in a coupled general circulation model: Unforced variations versus forced changes. *Journal Of Climate*, **18** (16), 3270–3293, doi:{10.1175/JCLI3481.1}.

Danabasoglu, G., 2008: On Multidecadal Variability of the Atlantic Meridional Overturning Circulation in the Community Climate System Model Version 3. *Journal Of Climate*, **21** (21), 5524–5544, doi:{10.1175/2008JCLI2019.1}.

Delworth, T., S. Manabe, and R. Stouffer, 1993: Interdecadal variations of the thermohaline circulation in a coupled ocean-atmosphere model. *Journal Of Climate*, **6** (11), 1993–2011, doi:{10.1175/1520-0442(1993)006\textless1993:IVOTTC\textgreater2.0.CO;2}.

Deser, C., J. E. Walsh, and M. S. Timlin, 2000: Arctic sea ice variability in the context of recent atmospheric circulation trends. *J. Climate*, **13** (3), 617–633, doi:doi:10.1175/1520-0442(2000)013\textless0617:ASIVIT\textgreater2.0.CO;2, URL [http://dx.doi.org/10.1175/1520-0442\(2000\)013\textless0617:ASI%VIT\textgreater2.0.CO;2](http://dx.doi.org/10.1175/1520-0442(2000)013\textless0617:ASI%VIT\textgreater2.0.CO;2).

Deshayes, J., R. Curry, and R. Msadek, 2014: Cmp5 model intercomparison of freshwater budget and circulation in the north atlantic. *Journal of Climate*, **27** (9), 3298–3317.

775 Dong, B., and R. Sutton, 2001: The dominant mechanisms of variability in Atlantic ocean heat
 776 transport in a coupled ocean-atmosphere GCM. *Geophysical Research Letters*, **28 (12)**, 2445–
 777 2448, doi:{10.1029/2000GL012531}.

778 Dong, B., and R. Sutton, 2005: Mechanism of interdecadal thermohaline circulation variability in
 779 a coupled ocean-atmosphere GCM. *Journal Of Climate*, **18 (8)**, 1117–1135.

780 Dunstone, N., D. Smith, and R. Eade, 2011: Multi-year predictability of the tropical atlantic atmo-
 781 sphere driven by the high latitude north atlantic ocean. *Geophysical Research Letters*, **38 (14)**.

782 Eden, C., and R. J. Greatbatch, 2003: A damped decadal oscillation in the north atlantic climate
 783 system. *Journal of climate*, **16 (24)**, 4043–4060.

784 Eden, C., and J. Willebrand, 2001: Mechanism of interannual to decadal variability of the North
 785 Atlantic circulation. *Journal Of Climate*, **14 (10)**, 2266–2280, doi:{10.1175/1520-0442(2001)
 786 014\textless2266:MOITDV\textgreater2.0.CO;2}.

787 Escudier, R., J. Mignot, and D. Swingedouw, 2013: A 20-year coupled ocean-sea ice-atmosphere
 788 variability mode in the north atlantic in an AOGCM. *Climate dynamics*, **40 (3-4)**, 619–636.

789 Fevrier, S., J. Sirven, and C. Herbaut, 2007: Interaction of a coastal kelvin wave with the mean
 790 state in the gulf stream separation area. *Journal Of Physical Oceanography*, **37 (6)**, 1429–1444,
 791 doi:{10.1175/JPO3062.1}.

792 Folland, C., T. Palmer, and D. Parker, 1986: Sahel rainfall and worldwide sea temperatures, 1901–
 793 85. *Nature*, **320 (6063)**, 602–607.

794 Frankcombe, L. M., A. von der Heydt, and H. A. Dijkstra, 2010: North atlantic multidecadal
 795 climate variability: An investigation of dominant time scales and processes. *Journal Of Climate*,
 796 **23 (13)**, 3626–3638, doi:{10.1175/2010JCLI3471.1}.

- 797 Frankignoul, C., P. Muller, and E. Zorita, 1997: A simple model of the decadal response of the
798 ocean to stochastic wind forcing. *Journal Of Physical Oceanography*, **27 (8)**, 1533–1546, doi:
799 {10.1175/1520-0485(1997)027\textless1533:ASMOTD\textgreater2.0.CO;2}.
- 800 Gelderloos, R., C. A. Katsman, and S. S. Drijfhout, 2011: Assessing the roles of three eddy
801 types in restratifying the labrador sea after deep convection. *Journal of Physical Oceanography*,
802 **41 (11)**, 2102–2119.
- 803 Goldenberg, S. B., C. W. Landsea, A. M. Mestas-Nuez, and W. M. Gray, 2001: The recent increase
804 in atlantic hurricane activity: Causes and implications. *Science*, **293 (5529)**, 474–479, doi:10.
805 1126/science.1060040, URL <http://www.sciencemag.org/content/293/5529/474.abstract>, [http://](http://www.sciencemag.org/content/293/5529/474.full.pdf)
806 www.sciencemag.org/content/293/5529/474.full.pdf.
- 807 Good, S. A., M. J. Martin, and N. A. Rayner, 2013: En4: Quality controlled ocean tempera-
808 ture and salinity profiles and monthly objective analyses with uncertainty estimates. *Journal of*
809 *Geophysical Research: Oceans*, **118 (12)**, 6704–6716.
- 810 Grotzner, A., M. Latif, and T. Barnett, 1998: A decadal climate cycle in the north atlantic ocean
811 as simulated by the ECHO coupled GCM. *Journal Of Climate*, **11 (5)**, 831–847, doi:{10.1175/
812 1520-0442(1998)011\textless0831:ADCCIT\textgreater2.0.CO;2}.
- 813 Hakkinen, S., and P. B. Rhines, 2004: Decline of subpolar north atlantic circulation during
814 the 1990s. *Science*, **304 (5670)**, 555–559, doi:10.1126/science.1094917, URL [http://www.](http://www.sciencemag.org/content/304/5670/555.abstract)
815 [sciencemag.org/content/304/5670/555.abstract](http://www.sciencemag.org/content/304/5670/555.abstract), [http://www.sciencemag.org/content/304/5670/](http://www.sciencemag.org/content/304/5670/555.full.pdf)
816 [555.full.pdf](http://www.sciencemag.org/content/304/5670/555.full.pdf).
- 817 Hodson, D. L., and R. T. Sutton, 2012: The impact of resolution on the adjustment and decadal
818 variability of the atlantic meridional overturning circulation in a coupled climate model. *Climate*

dynamics, **39** (12), 3057–3073.

Holland, M., C. Bitz, M. Eby, and A. Weaver, 2001: The role of ice-ocean interactions in the variability of the North Atlantic thermohaline circulation. *Journal Of Climate*, **14** (5), 656–675, doi:{10.1175/1520-0442(2001)014\textless0656:TROI\textgreater2.0.CO;2}.

Ingleby, B., and M. Huddleston, 2007: Quality control of ocean temperature and salinity profiles. historical and real-time data. *Journal of Marine Systems*, **65** (1), 158–175.

Johnson, H., and D. Marshall, 2002: Localization of abrupt change in the North Atlantic thermohaline circulation. *Geophysical Research Letters*, **29** (6), doi:{10.1029/2001GL014140}.

Jungclauss, J., H. Haak, M. Latif, and U. Mikolajewicz, 2005: Arctic-North Atlantic interactions and multidecadal variability of the meridional overturning circulation. *Journal Of Climate*, **18** (19), 4013–4031.

Kara, A., P. Rochford, and H. Hurlburt, 2000: An optimal definition for ocean mixed layer depth. *Journal Of Geophysical Research-Oceans*, **105** (C7), 16 803–16 821, doi:{10.1029/2000JC900072}.

Kleppin, H., M. Jochum, B. Otto-Bliesner, C. A. Shields, and S. Yeager, 2015: Stochastic atmospheric forcing as trigger for sudden greenland warmings. *Journal of Climate* (submitted).

Knight, J., R. Allan, C. Folland, M. Vellinga, and M. Mann, 2005: A signature of persistent natural thermohaline circulation cycles in observed climate. *Geophysical Research Letters*, **32** (20), doi:{10.1029/2005GL024233}.

Kwon, Y.-O., M. A. Alexander, N. A. Bond, C. Frankignoul, H. Nakamura, B. Qiu, and L. A. Thompson, 2010: Role of the gulf stream and kuroshio-oyashio systems in large-scale atmosphere-ocean interaction: A review. *Journal of Climate*, **23** (12), 3249–3281.

841 Kwon, Y.-O., and C. Frankignoul, 2014: Mechanisms of multidecadal atlantic meridional over-
842 turning circulation variability diagnosed in depth versus density space. *Journal of Climate*,
843 **27 (24)**, 9359–9376.

844 MacLachlan, C., and Coauthors, 2014: Global seasonal forecast system version 5 (glosea5): a high
845 resolution seasonal forecast system. *Quarterly Journal of the Royal Meteorological Society*.

846 MacMartin, D. G., E. Tziperman, and L. Zanna, 2013: Frequency domain multimodel analysis
847 of the response of atlantic meridional overturning circulation to surface forcing. *Journal of*
848 *Climate*, **26 (21)**, 8323–8340.

849 Madec, G., and Coauthors, 2008: Nemo ocean engine: Note du pole de modélisation, institut
850 pierre-simon laplace (ipsl), france, no 27 issn no 1288-1619, available at: [http://www.nemo-](http://www.nemo-ocean.eu)
851 [ocean.eu](http://www.nemo-ocean.eu). Tech. rep., IPSL LSCE, UVSQ, CEA CNRS, Unite Mixte, Bat 712, F-91191 Gif Sur
852 Yvette, France.

853 Mann, M., J. Park, and R. Bradley, 1995: Global Interdecadal And Century-Scale Climate Oscil-
854 lations During The Past 5 Centuries. *Nature*, **378 (6554)**, 266–270, doi:{10.1038/378266a0}.

855 Marzocchi, A., J. J.-M. Hirschi, N. P. Holliday, S. A. Cunningham, A. T. Blaker, and A. C. Coward,
856 2015: The north atlantic subpolar circulation in an eddy-resolving global ocean model. *Journal*
857 *of Marine Systems*, **142 (0)**, 126 – 143, doi:<http://dx.doi.org/10.1016/j.jmarsys.2014.10.007>,
858 URL <http://www.sciencedirect.com/science/article/pii/S09247963140%02437>.

859 Matsueda, M., R. Mizuta, and S. Kusunoki, 2009: Future change in wintertime atmospheric block-
860 ing simulated using a 20-km-mesh atmospheric global circulation model. *Journal of Geophysi-*
861 *cal Research: Atmospheres (1984–2012)*, **114 (D12)**.

- 862 Megann, A., and Coauthors, 2014: Go5. 0: the joint nerc–met office nemo global ocean model for
863 use in coupled and forced applications. *Geoscientific Model Development*, **7** (3), 1069–1092.
- 864 Menary, M. B., D. L. R. Hodson, J. I. Robson, R. T. Sutton, R. A. Wood, and J. A. Hunt,
865 2015: Exploring the impact of cmip5 model biases on the simulation of north atlantic decadal
866 variability. *Geophysical Research Letters*, n/a–n/a, doi:10.1002/2015GL064360, URL <http://dx.doi.org/10.1002/2015GL064360>, 2015GL064360.
- 867
- 868 Menary, M. B., W. Park, K. Lohmann, M. Vellinga, M. D. Palmer, M. Latif, and J. H. Jungclaus,
869 2012: A multimodel comparison of centennial Atlantic meridional overturning circulation vari-
870 ability. *Climate Dynamics*, **38** (11-12), 2377–2388, doi:{ 10.1007/s00382-011-1172-4}.
- 871 Nonaka, M., and S.-P. Xie, 2003: Covariations of sea surface temperature and wind
872 over the kuroshio and its extension: Evidence for ocean-to-atmosphere feedback*. *J.*
873 *Climate*, **16** (9), 1404–1413, doi:doi:10.1175/1520-0442(2003)16\textless1404:COSSA\textgreater2.0.CO;2, URL [http://dx.doi.org/10.1175/1520-0442\(2003\)16\textless1404:COSSA\textgreater2.0.CO;2](http://dx.doi.org/10.1175/1520-0442(2003)16\textless1404:COSSA\textgreater2.0.CO;2).
- 874 TA\textgreater2.0.CO;2.
- 875
- 876 Polo, I., J. Robson, R. Sutton, and M. A. Balmaseda, 2014: The importance of wind and buoyancy
877 forcing for the boundary density variations and the geostrophic component of the amoc at 26 n.
878 *Journal of Physical Oceanography*, **44** (9), 2387–2408.
- 879 Roberts, C., and Coauthors, 2013: Atmosphere drives recent interannual variability of the atlantic
880 meridional overturning circulation at 26.5 n. *Geophysical Research Letters*, **40** (19), 5164–5170.
- 881 Robson, J. I., 2010: Understanding the performance of a decadal prediction system. Ph.D. thesis,
882 The University of Reading, Reading, UK.

- 883 Rodwell, M., D. Rowell, and C. Folland, 1999: Oceanic forcing of the wintertime North Atlantic
884 Oscillation and European climate. *Nature*, **398 (6725)**, 320–323, doi:{10.1038/18648}.
- 885 Roussenov, V. M., R. G. Williams, C. W. Hughes, and R. J. Bingham, 2008: Boundary wave
886 communication of bottom pressure and overturning changes for the North Atlantic. *Journal Of*
887 *Geophysical Research-Oceans*, **113 (C8)**, doi:{10.1029/2007JC004501}.
- 888 Sarafanov, A., A. Falina, A. Sokov, and A. Demidov, 2008: Intense warming and salinification
889 of intermediate waters of southern origin in the eastern subpolar north atlantic in the 1990s
890 to mid-2000s. *Journal of Geophysical Research: Oceans*, **113 (C12)**, n/a–n/a, doi:10.1029/
891 2008JC004975, URL <http://dx.doi.org/10.1029/2008JC004975>.
- 892 Saravanan, R., and J. C. McWilliams, 1998: Advective ocean-atmosphere interaction: An ana-
893 lytical stochastic model with implications for decadal variability. *Journal of Climate*, **11 (2)**,
894 165–188.
- 895 Scaife, A. A., and Coauthors, 2011: Improved atlantic winter blocking in a climate model. *Geo-*
896 *physical Research Letters*, **38 (23)**.
- 897 Schauer, U., and A. Beszczynska-Möller, 2009: Problems with estimation and interpretation of
898 oceanic heat transport–conceptual remarks for the case of fram strait in the arctic ocean. *Ocean*
899 *Science*, **5 (4)**, 487–494.
- 900 Screen, J. A., 2013: Influence of arctic sea ice on european summer precipitation. *Environmental*
901 *Research Letters*, **8 (4)**, 044 015.
- 902 Sévellec, F., and A. V. Fedorov, 2013: The leading, interdecadal eigenmode of the atlantic merid-
903 ional overturning circulation in a realistic ocean model. *Journal of Climate*, **26 (7)**, 2160–2183.

904 Sicre, M.-A., and Coauthors, 2008: A 4500-year reconstruction of sea surface temperature vari-
 905 ability at decadal time-scales off North Iceland. *Quaternary Science Reviews*, **27** (21-22), 2041–
 906 2047, doi:{10.1016/j.quascirev.2008.08.009}.

907 Smith, D. M., R. Eade, N. J. Dunstone, D. Fereday, J. M. Murphy, H. Pohlmann, and A. A. Scaife,
 908 2010: Skilful multi-year predictions of atlantic hurricane frequency. *Nature Geoscience*, **3** (12),
 909 846–849.

910 Spence, P., O. A. Saenko, W. Sijp, and M. England, 2011: The role of bottom pressure torques
 911 on the interior pathways of north atlantic deep water. *J. Phys. Oceanogr.*, **42** (1), 110–125,
 912 doi:doi:10.1175/2011JPO4584.1, URL <http://dx.doi.org/10.1175/2011JPO4584.1>.

913 Sutton, R., and D. Hodson, 2005: Atlantic Ocean forcing of North American and European sum-
 914 mer climate. *Science*, **309** (5731), 115–118, doi:{10.1126/science.1109496}.

915 Timmermann, A., M. Latif, R. Voss, and A. Grotzner, 1998: Northern Hemispheric interdecadal
 916 variability: A coupled air-sea mode. *Journal Of Climate*, **11** (8), 1906–1931, doi:{10.1175/
 917 1520-0442-11.8.1906}.

918 Tréguier, A.-M., S. Theetten, E. P. Chassignet, T. Penduff, R. Smith, L. Talley, J. Beismann, and
 919 C. Böning, 2005: The north atlantic subpolar gyre in four high-resolution models. *Journal of*
 920 *Physical Oceanography*, **35** (5), 757–774.

921 Treguier, A. M., and Coauthors, 2014: Meridional transport of salt in the global ocean from an
 922 eddy-resolving model. *Ocean Science*, **10** (2), 243–255, doi:10.5194/os-10-243-2014, URL
 923 <http://www.ocean-sci.net/10/243/2014/>.

924 Tulloch, R., and J. Marshall, 2012: Exploring mechanisms of variability and predictability of
 925 atlantic meridional overturning circulation in two coupled climate models. *Journal of Climate*,
 926 **25 (12)**, 4067–4080.

927 Vellinga, M., and P. Wu, 2004: Low-latitude freshwater influence on centennial variability of the
 928 Atlantic thermohaline circulation. *Journal Of Climate*, **17 (23)**, 4498–4511.

929 Visbeck, M., H. Cullen, G. Krahmann, and N. Naik, 1998: An ocean model’s response to North
 930 Atlantic Oscillation-like wind forcing. *Geophysical Research Letters*, **25 (24)**, 4521–4524, doi:
 931 {10.1029/1998GL900162}.

932 Volkov, D. L., T. Lee, and L.-L. Fu, 2008: Eddy-induced meridional heat transport in the ocean.
 933 *Geophysical Research Letters*, **35 (20)**, n/a–n/a, doi:10.1029/2008GL035490, URL [http://dx.](http://dx.doi.org/10.1029/2008GL035490)
 934 [doi.org/10.1029/2008GL035490](http://dx.doi.org/10.1029/2008GL035490).

935 Walters, D., and Coauthors, 2011: The Met Office Unified Model global atmosphere 3.0/3.1 and
 936 JULES global land 3.0/3.1 configurations. *Geoscientific Model Development Discussions*, **4 (2)**,
 937 1213–1271.

938 Wang, C., L. Zhang, S.-K. Lee, L. Wu, and C. R. Mechoso, 2014: A global perspective on cmip5
 939 climate model biases. *Nature Climate Change*, **4 (3)**, 201–205, doi:10.1038/nclimate2118, URL
 940 <http://dx.doi.org/10.1038/nclimate2118>.

941 Watanabe, M., M. Kimoto, T. Nitta, and M. Kachi, 1999: A comparison of decadal climate oscil-
 942 lations in the north Atlantic detected in observations and a coupled GCM. *Journal Of Climate*,
 943 **12 (9)**, 2920–2940, doi:{10.1175/1520-0442(1999)012\textless2920:ACODCO\textgreater2.
 944 0.CO;2}.

945 Wilks, D., 1997: Resampling hypothesis tests for autocorrelated fields. *Journal Of Climate*, **10** (1),
 946 65–82.

947 Wohllleben, T., and A. Weaver, 1995: Interdecadal Climate Variability In The Subpolar North-
 948 Atlantic. *Climate Dynamics*, **11** (8), 459–467, doi:{10.1007/s003820050088}.

949 Yeager, S., 2015: Topographic coupling of the atlantic overturning and gyre circula-
 950 tions. *J. Phys. Oceanogr.*, doi:doi:10.1175/JPO-D-14-0100.1, URL [http://dx.doi.org/10.1175/](http://dx.doi.org/10.1175/JPO-D-14-0100.1)
 951 [JPO-D-14-0100.1](http://dx.doi.org/10.1175/JPO-D-14-0100.1).

952 Yeager, S., and G. Danabasoglu, 2012: Sensitivity of atlantic meridional overturning circulation
 953 variability to parameterized nordic sea overflows in ccsm4. *Journal of Climate*, **25** (6), 2077–
 954 2103.

955 Zhang, R., 2010: Latitudinal dependence of Atlantic meridional overturning circulation (AMOC)
 956 variations. *Geophysical Research Letters*, **37**, doi:{10.1029/2010GL044474}.

957 Zhang, R., and T. L. Delworth, 2006: Impact of Atlantic multidecadal oscillations on In-
 958 dia/Sahel rainfall and Atlantic hurricanes. *Geophysical Research Letters*, **33** (17), doi:{10.1029/
 959 2006GL026267}.

960 Zhang, R., and G. K. Vallis, 2007: The role of bottom vortex stretching on the path of the north
 961 Atlantic western boundary current and on the northern recirculation gyre. *Journal Of Physical*
 962 *Oceanography*, **37** (8), 2053–2080, doi:{10.1175/JPO3102.1}.

963	LIST OF TABLES	
964	Table 1.	Time mean simulated heat fluxes into the North Atlantic Current (NAC) and
965		Irminger Current regions (TW, referenced to 0°C). 46
966	Table 2.	Correlations (regression slopes in brackets) between net ocean heat transport
967		(νT) and advective heat flux components in the North Atlantic Current (NAC)
968		and Irminger Current at various timescales (TW). The 95% significance levels,
969		assuming a two-tailed t-test and accounting for some missing data, are 0.03,
970		0.12, and 0.37 for monthly, annual, and decadal data respectively. 47
971	Table 3.	Standard deviations of advective heat flux components in the North Atlantic
972		Current (NAC) and Irminger Current at various timescales (TW). 48
973	Table 4.	Characteristic magnitudes of density changes (kg/m^3) in different simu-
974		lated/estimated temperature or salinity (T/S) regimes. Mean states are volume
975		averaged temperature and salinity (in the models defined as the observed mean
976		plus a model bias, <i>e.g.</i> EN4 + HG3 bias) in the Irminger Current (43–45°W,
977		58–60°N, top 500m). The density changes are calculated by estimating the
978		decadal standard deviation (s.d.) in temperature or salinity (by band-pass filter-
979		ing the data to allow only periods in the range 10–30 years) and recalculating
980		the densities with these T/S perturbations added. As there is limited raw data
981		from EN4 to reliably estimate decadal variability in the Irminger Current, and
982		to simplify the experimental design and interpretation, we use HG3 estimates
983		of the decadal variability in temperature and salinity in all cases. 49

	NAC	Irminger Current
East advection	-1660	493
West advection	1753	-392
North advection	33	0
South advection	-83.1	-23.7
Net vertical advection	7.5	-63.9
Net convergence	51.4	15.2
Surface	-49.8	-16.9
A: Sum of advection and surface (net sum)	1.6	-1.7
B: Ocean heat content change (net actual)	0.1	0.1
Correlation A:B (Monthly, Annual, Decadal)	0.96, 0.93, 0.98	0.96, 0.94, 0.95

984 TABLE 1. Time mean simulated heat fluxes into the North Atlantic Current (NAC) and Irminger Current
985 regions (TW, referenced to 0°C).

	Monthly	Annual	Decadal
NAC $v_0 T'$	0.00 (0.01)	-0.16 (-1.1)	-0.23 (-1.6)
NAC $v' T_0$	0.29 (0.82)	0.36 (2.4)	0.42 (2.9)
Irminger Current $v_0 T'$	0.83 (0.83)	0.34 (0.66)	0.29 (0.53)
Irminger Current $v' T_0$	0.19 (0.10)	-0.10 (-0.18)	-0.14 (-0.24)

986 TABLE 2. Correlations (regression slopes in brackets) between net ocean heat transport (vT) and advective
 987 heat flux components in the North Atlantic Current (NAC) and Irminger Current at various timescales (TW).
 988 The 95% significance levels, assuming a two-tailed t-test and accounting for some missing data, are 0.03, 0.12,
 989 and 0.37 for monthly, annual, and decadal data respectively.

	Monthly	Annual	Decadal
NAC $v_0 T'$	139	43	31
NAC $v' T_0$	149	44	33
NAC $v' T'$	58	16	11
Irminger Current $v_0 T'$	13.1	4.0	3.2
Irminger Current $v' T_0$	6.7	3.6	2.7
Irminger Current $v' T'$	4.0	1.1	1.0

990 TABLE 3. Standard deviations of advective heat flux components in the North Atlantic Current (NAC) and
991 Irminger Current at various timescales (TW).

Mean state	Density change for one s.d. change in T	Density change for one s.d. change in S
EN4 + HG3 bias	0.027	0.014
EN4 + IPSL bias	0.010	0.014
EN4 (original)	0.023	0.014

992 TABLE 4. Characteristic magnitudes of density changes (kg/m^3) in different simulated/estimated temperature
 993 or salinity (T/S) regimes. Mean states are volume averaged temperature and salinity (in the models defined as
 994 the observed mean plus a model bias, *e.g.* EN4 + HG3 bias) in the Irminger Current (43–45°W, 58–60°N, top
 995 500m). The density changes are calculated by estimating the decadal standard deviation (s.d.) in temperature
 996 or salinity (by band-pass filtering the data to allow only periods in the range 10–30 years) and recalculating
 997 the densities with these T/S perturbations added. As there is limited raw data from EN4 to reliably estimate
 998 decadal variability in the Irminger Current, and to simplify the experimental design and interpretation, we use
 999 HG3 estimates of the decadal variability in temperature and salinity in all cases.

LIST OF FIGURES

- 1001 **Fig. 1.** A summary of some of the literature on simulated decadal variability in the North Atlantic
 1002 subpolar gyre (NA SPG), with a particular emphasis on studies which found self-sustaining
 1003 cyclical behaviour. Key regions of the NA SPG are marked. The figure legend (right)
 1004 denotes the studies which we have attempted to synthesise and an associated numerical
 1005 identifier. Where these studies report a significant peak in the power spectrum on decadal
 1006 timescales this is noted as well as whether the mechanism is primarily ocean-only or inher-
 1007 ently coupled. Studies where the atmosphere is postulated to amplify — but not transmute
 1008 — the signal are marked with an asterisk. For each study the feedback or process which
 1009 is reported as crucial in setting the timescale is marked on the map using a simple number-
 1010 ing system. These comprise: 1) Feedbacks relating to the deep water pathways and their
 1011 interaction with the northward flowing western boundary current, 2) Rossby wave (or some-
 1012 times ‘geostrophic self advection’) transit times across the NA SPG, 3) the mean advection
 1013 timescale for anomalies to propagate into the NA SPG from the tropics, or for small anoma-
 1014 lies to integrate up over time. Lastly, using the same numerical key, the studies are split into
 1015 which of temperature or salinity is reported to control decadal timescale density changes in
 1016 the Labrador Sea. In all case, studies in brackets appear in more than one category. This
 1017 represents a drastic simplification of each of these studies and the reader is referred to the
 1018 original works for further details. In particular, the reported feedback/process that sets the
 1019 overall timescale to some degree also reflects the precise focus of the particular study. . . . 53
- 1020 **Fig. 2.** Top 500m depth averaged temperature (T500, a) salinity (S500, b), and density (ρ 500, c)
 1021 biases in HG3 (computed from full time series) compared to EN4. Grey shading is used
 1022 for regions shallower than 500m. d) Standard deviation in March mixed layer depths (Kara
 1023 et al. 2000), to highlight where deep convection occurs 54
- 1024 **Fig. 3.** a) Time mean Atlantic overturning streamfunction in HG3. The contour interval is 2Sv and
 1025 the zero-line is marked with a grey contour. At 26.5°N the profile from the RAPID array is
 1026 overlaid on the same colour/contour scale. The depth of the maximum in the RAPID profile
 1027 is marked with a cross. Note that the latitudes north of 45°N are approximate (within 1°)
 1028 due to the increasingly curved nature of the model grid towards the two northern poles. b)
 1029 Time mean NA SPG barotropic streamfunction in HG3. Contour interval is 10Sv. c) Time
 1030 series of the overturning streamfunction at 26.5°N and 1000m in HG3 (red). Also shown
 1031 are the time mean and annual mean standard deviation from the 10 years of RAPID data
 1032 (black). d) Time series of the minimum (multiplied by -1) of the barotropic streamfunction
 1033 in the NA SPG in HG3. 55
- 1034 **Fig. 4.** a) Time mean top 500m depth averaged temperature (T500) in HG3. Contours at 6°C and
 1035 10°C are also marked (black) to show the shape of the gyre and for comparison with equiv-
 1036 alent contours from EN4 (grey). Areas in white are shallower than 500m. The dashed grey
 1037 box denotes the four quadrants and fifth overall region for which power spectra of T500
 1038 were produced. The Irminger Current region (red box) and Gulf Stream/North Atlantic Cur-
 1039 rent region (blue box) analysed in the text are also marked. The dashed black line stretching
 1040 south from the Grand Banks denotes the transect location for dynamic height analysis. b)
 1041 The T500 power spectrum for the whole subpolar gyre region (combination of all four quad-
 1042 rants). An estimate of significance is given by the 5–95% confidence intervals for a red noise
 1043 process with the same mean and standard deviation. Periods of 16–17 years are highlighted
 1044 with the blue shading. c) As (b) but for the NAO index, defined as the difference between
 1045 simulated sea level pressures over the Azores and Iceland. 56
- 1046 **Fig. 5.** Regressions between basin-wide North Atlantic (45–65°N) sea surface temperatures (SSTs)
 1047 and, from left to right: SST, top 500m depth averaged temperature (T500), net surface heat

1048	flux into ocean (SHF), sea surface salinity (SSS), wintertime mean sea level pressure (DJF	
1049	MSLP), and ice fraction. From top to bottom, the SST index lags then leads the fields from	
1050	-6 to +4 and then +11 years. The same colour palette is used for each regression map with	
1051	the units and scale for each regression slope shown below.	57
1052	Fig. 6. The lagged correlation between the Irminger Current and North Atlantic Current top 500m	
1053	depth averaged temperatures (T500). Regions are as marked in Figure 4a. Time series have	
1054	been filtered with a band-pass filter of 5–150 years to highlight the decadal correlations	
1055	by removing annual variability and the long term drift. An estimate of the significance is	
1056	provided by the 95% (red) and 99% (blue) confidence intervals estimated by creating 40,000	
1057	random time series with the same mean, standard deviation, and applied filtering.	58
1058	Fig. 7. a) 5-year smoothed Labrador Sea (50–60°W, 55–62°N) top 500m mean density (black) and	
1059	contributions from temperature (by keeping salinity at the time mean in the density equation	
1060	of state, blue) and from salinity (by keeping temperature at the time mean in the density	
1061	equation of state, red). Temperature and salinity both linearly detrended prior to computing	
1062	density. b) Lagged correlation of temperature-induced (blue) and salinity-induced (red) den-	
1063	sity against actual density for detrended data, either unsmoothed (dashed) or pre-smoothed	
1064	with a 5-year running mean (solid, as in (a)). Temperature/salinity-induced density leads	
1065	actual density at negative lags.	59
1066	Fig. 8. Transect south from the Grand Banks through the North Atlantic Current at 47.5°W, as	
1067	shown in Figure 4a. Density (referenced to 0m) composite of high minus low densities in the	
1068	Labrador Sea, computed by averaging all cases where Labrador Sea volume mean density	
1069	(computed over the region 47–55°W, 56–61°, 0–1000m) was at least one standard deviation	
1070	larger than the time mean, and subtracting the average of all cases where density was at least	
1071	one standard deviation less than the time mean. Cross sections lag by 1 year the Labrador	
1072	Sea index. b) As (a) but for dynamic height composites (relative to 1500m). c) As (a) but for	
1073	the geostrophic circulation (relative to 1500m). d) As (c) but first removing the NAO signal	
1074	from the density field after Polo et al. (2014), see text. Cross-sectional time mean density	
1075	is indicated by the black contours. All data has been detrended and 5-year smoothed. Only	
1076	data significant at the 99% level is shown. Significance is estimated using a moving blocks	
1077	bootstrap approach (Wilks 1997) reconstructing the composites 10000 times by resampling	
1078	the data with a block length estimated from the autocorrelation in each original composite.	60
1079	Fig. 9. A schematic of the proposed mechanism. The various processes in different regions,	
1080	the timescales, and the postulated role of the atmosphere are as described in the text.	
1081	Dashed grey lines denote the approximate location of NA SPG and subtropical gyres with	
1082	bathymetry of particular interest marked brown. Also marked are regions where ocean heat	
1083	transport correlates with circulation anomalies (orange) or is dominated by temperature	
1084	anomalies (red) and where the negative feedback is suggested to occur (blue). Addition-	
1085	ally, black dashed lines denote regions where the atmosphere is postulated to play a role in	
1086	forcing or feeding back on ocean anomalies.	61
1087	Fig. 10. Mean sea level pressure (MSLP) composites created using area-averaged eastern NA SPG	
1088	SSTs (10–30°W, 50–65°) to highlight the asymmetry between positive and negative phases	
1089	of the proposed mechanism. a) Composite created using the highest 10% of SST anom-	
1090	alies. b) As (a) but for the lowest 10% of SST anomalies. Only data significant at the 99%	
1091	level is shown. Significance is estimated using a moving blocks bootstrap approach (Wilks	
1092	1997) reconstructing the composites 10000 times by resampling the data with a block length	
1093	estimated from the autocorrelation in each original composite.	62

1094 **Fig. 11.** The full depth heat budget of the NA SPG (53–73°N) volume plotted using 9 year running
1095 means for clarity and to highlight decadal variability. Positive is into the specified region.
1096 a) Individual components of the heat budget as denoted in the legend. b) The anomalous
1097 heat budget (referenced against years 22–42) to highlight the trends in latent heat fluxes
1098 and advective heat fluxes through the southern boundary. Note that the heat content change
1099 ($\frac{dT}{dt}$, black) and sum of heat fluxes (red) do not match prior to the year 100 as instantaneous
1100 ocean temperatures (used to calculate $\frac{dT}{dt}$) were stored with intermittent frequency during
1101 this period. 63

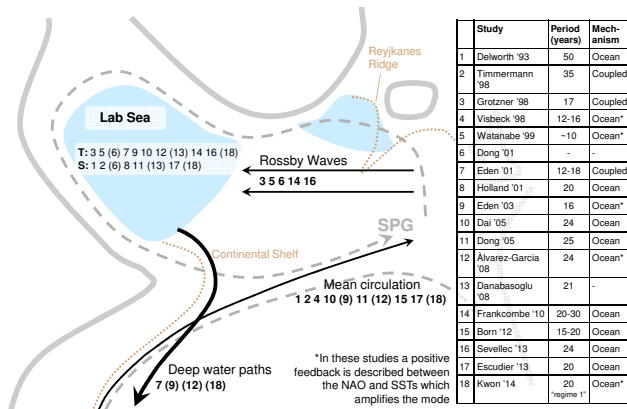


FIG. 1. A summary of some of the literature on simulated decadal variability in the North Atlantic subpolar gyre (NA SPG), with a particular emphasis on studies which found self-sustaining cyclical behaviour. Key regions of the NA SPG are marked. The figure legend (right) denotes the studies which we have attempted to synthesise and an associated numerical identifier. Where these studies report a significant peak in the power spectrum on decadal timescales this is noted as well as whether the mechanism is primarily ocean-only or inherently coupled. Studies where the atmosphere is postulated to amplify — but not transmute — the signal are marked with an asterisk. For each study the feedback or process which is reported as crucial in setting the timescale is marked on the map using a simple numbering system. These comprise: 1) Feedbacks relating to the deep water pathways and their interaction with the northward flowing western boundary current, 2) Rossby wave (or sometimes ‘geostrophic self advection’) transit times across the NA SPG, 3) the mean advection timescale for anomalies to propagate into the NA SPG from the tropics, or for small anomalies to integrate up over time. Lastly, using the same numerical key, the studies are split into which of temperature or salinity is reported to control decadal timescale density changes in the Labrador Sea. In all case, studies in brackets appear in more than one category. This represents a drastic simplification of each of these studies and the reader is referred to the original works for further details. In particular, the reported feedback/process that sets the overall timescale to some degree also reflects the precise focus of the particular study.

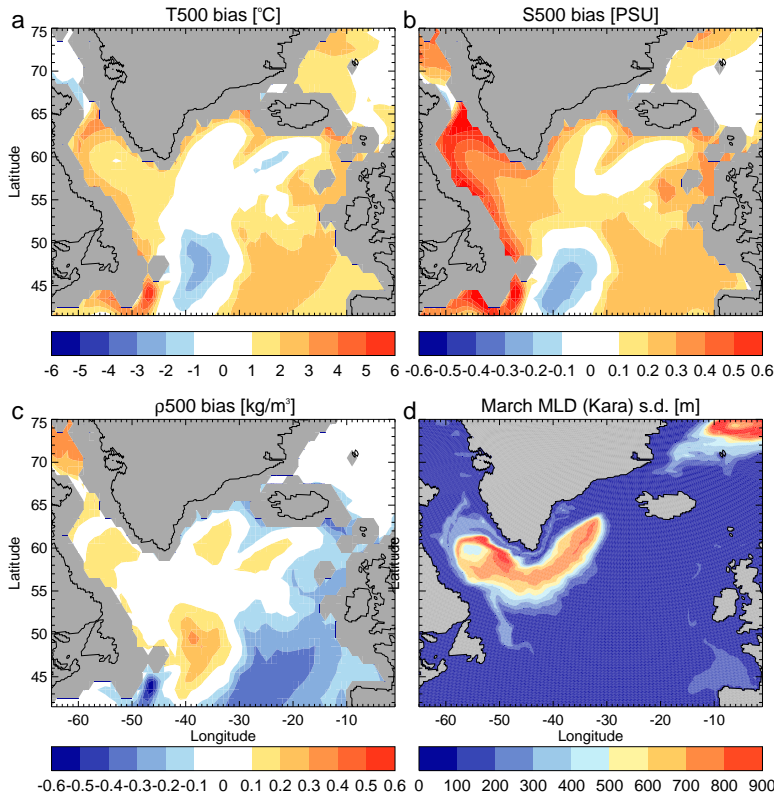


FIG. 2. Top 500m depth averaged temperature (T500, a) salinity (S500, b), and density (ρ 500, c) biases in
 HG3 (computed from full time series) compared to EN4. Grey shading is used for regions shallower than 500m.
 d) Standard deviation in March mixed layer depths (Kara et al. 2000), to highlight where deep convection occurs

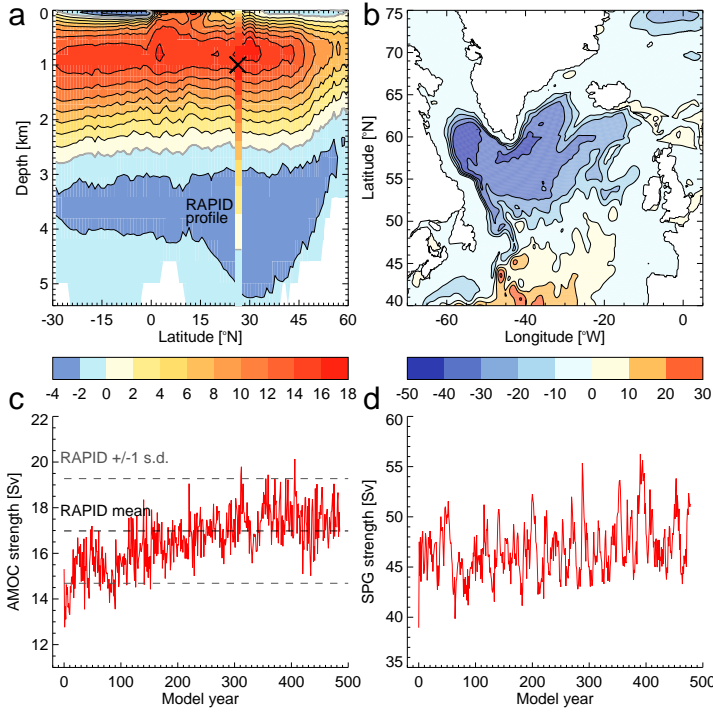


FIG. 3. a) Time mean Atlantic overturning streamfunction in HG3. The contour interval is 2Sv and the zero-line is marked with a grey contour. At 26.5°N the profile from the RAPID array is overlaid on the same colour/contour scale. The depth of the maximum in the RAPID profile is marked with a cross. Note that the latitudes north of 45°N are approximate (within 1°) due to the increasingly curved nature of the model grid towards the two northern poles. b) Time mean NA SPG barotropic streamfunction in HG3. Contour interval is 10Sv. c) Time series of the overturning streamfunction at 26.5°N and 1000m in HG3 (red). Also shown are the time mean and annual mean standard deviation from the 10 years of RAPID data (black). d) Time series of the minimum (multiplied by -1) of the barotropic streamfunction in the NA SPG in HG3.

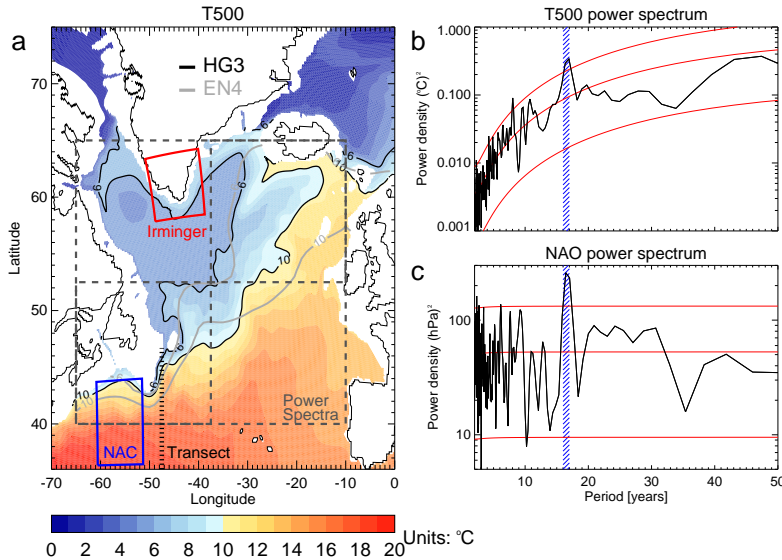


FIG. 4. a) Time mean top 500m depth averaged temperature (T500) in HG3. Contours at 6°C and 10°C are also marked (black) to show the shape of the gyre and for comparison with equivalent contours from EN4 (grey). Areas in white are shallower than 500m. The dashed grey box denotes the four quadrants and fifth overall region for which power spectra of T500 were produced. The Irminger Current region (red box) and Gulf Stream/North Atlantic Current region (blue box) analysed in the text are also marked. The dashed black line stretching south from the Grand Banks denotes the transect location for dynamic height analysis. b) The T500 power spectrum for the whole subpolar gyre region (combination of all four quadrants). An estimate of significance is given by the 5–95% confidence intervals for a red noise process with the same mean and standard deviation. Periods of 16–17 years are highlighted with the blue shading. c) As (b) but for the NAO index, defined as the difference between simulated sea level pressures over the Azores and Iceland.

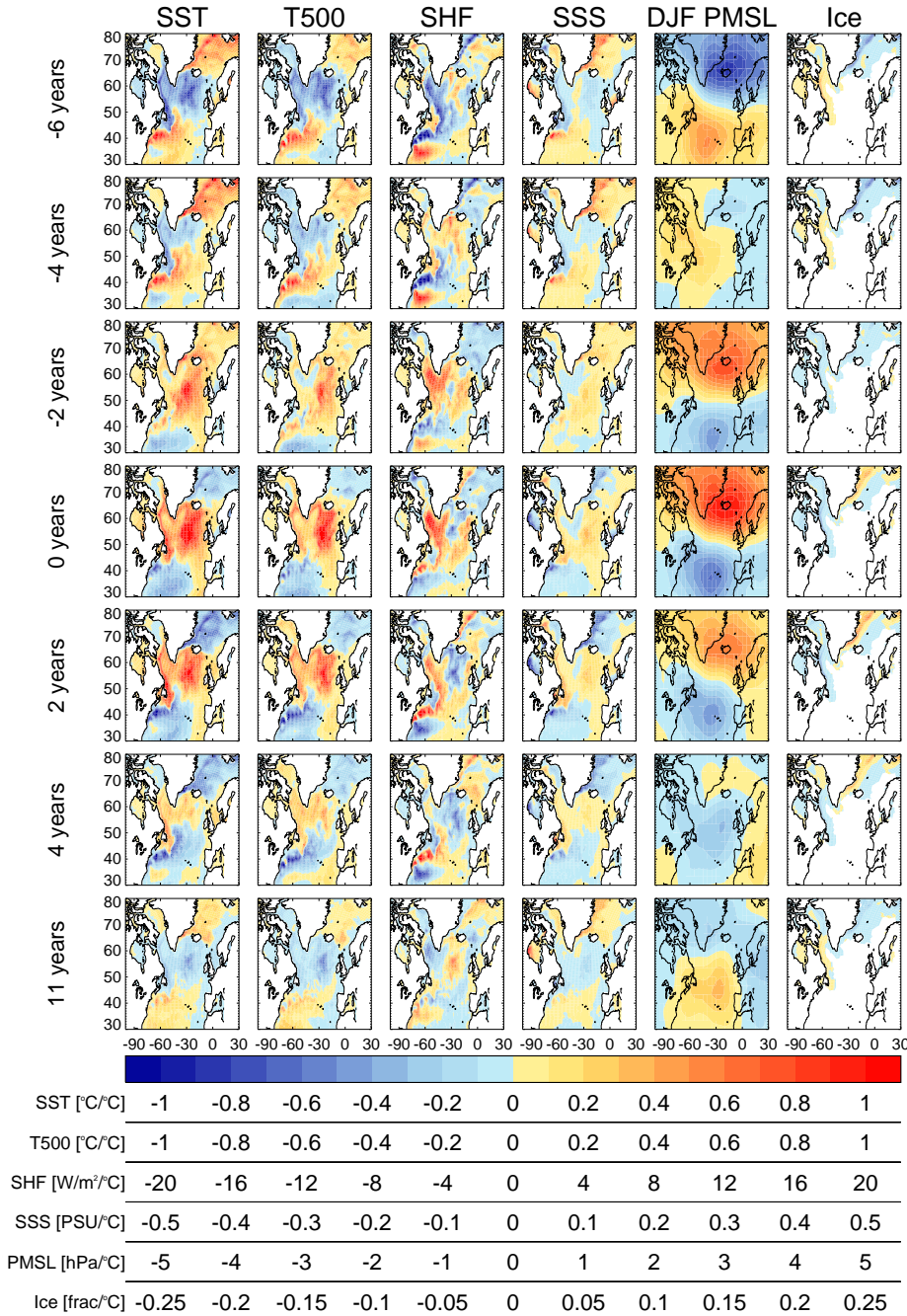


FIG. 5. Regressions between basin-wide North Atlantic (45–65°N) sea surface temperatures (SSTs) and, from left to right: SST, top 500m depth averaged temperature (T500), net surface heat flux into ocean (SHF), sea surface salinity (SSS), wintertime mean sea level pressure (DJF MSLP), and ice fraction. From top to bottom, the SST index lags then leads the fields from -6 to +4 and then +11 years. The same colour palette is used for each regression map with the units and scale for each regression slope shown below.

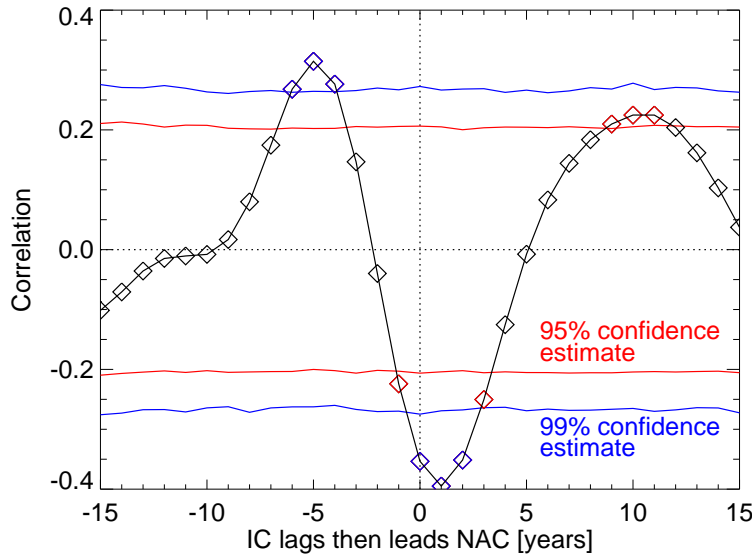


FIG. 6. The lagged correlation between the Irminger Current and North Atlantic Current top 500m depth averaged temperatures (T500). Regions are as marked in Figure 4a. Time series have been filtered with a band-pass filter of 5–150 years to highlight the decadal correlations by removing annual variability and the long term drift. An estimate of the significance is provided by the 95% (red) and 99% (blue) confidence intervals estimated by creating 40,000 random time series with the same mean, standard deviation, and applied filtering.

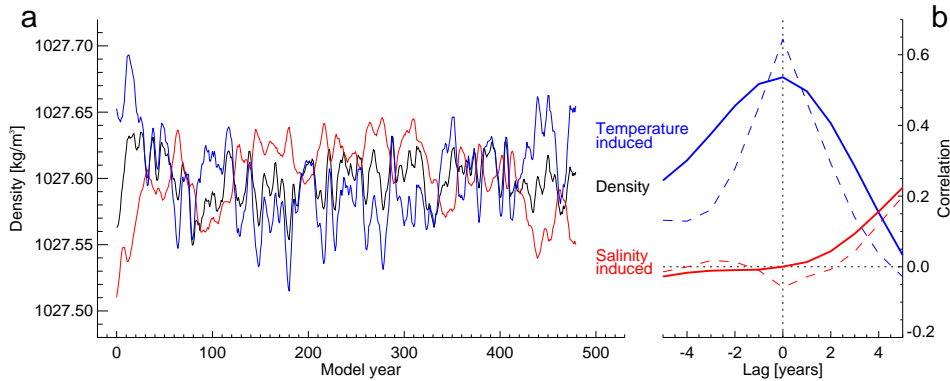


FIG. 7. a) 5-year smoothed Labrador Sea (50–60°W, 55–62°N) top 500m mean density (black) and contributions from temperature (by keeping salinity at the time mean in the density equation of state, blue) and from salinity (by keeping temperature at the time mean in the density equation of state, red). Temperature and salinity both linearly detrended prior to computing density. b) Lagged correlation of temperature-induced (blue) and salinity-induced (red) density against actual density for detrended data, either unsmoothed (dashed) or pre-smoothed with a 5-year running mean (solid, as in (a)). Temperature/salinity-induced density leads actual density at negative lags.

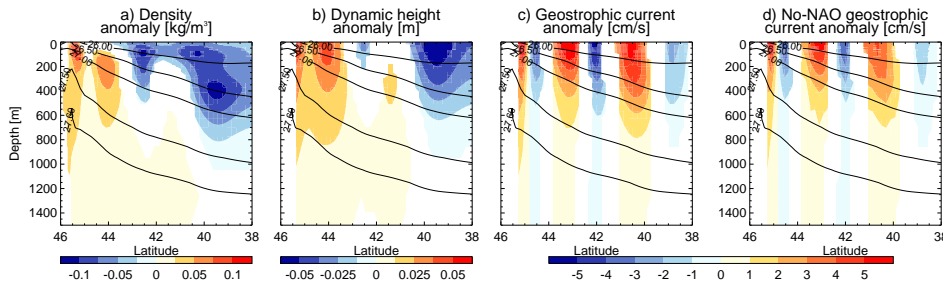


FIG. 8. Transect south from the Grand Banks through the North Atlantic Current at 47.5°W , as shown in Figure 4a. Density (referenced to 0m) composite of high minus low densities in the Labrador Sea, computed by averaging all cases where Labrador Sea volume mean density (computed over the region $47\text{--}55^{\circ}\text{W}$, $56\text{--}61^{\circ}$, $0\text{--}1000\text{m}$) was at least one standard deviation larger than the time mean, and subtracting the average of all cases where density was at least one standard deviation less than the time mean. Cross sections lag by 1 year the Labrador Sea index. b) As (a) but for dynamic height composites (relative to 1500m). c) As (a) but for the geostrophic circulation (relative to 1500m). d) As (c) but first removing the NAO signal from the density field after Polo et al. (2014), see text. Cross-sectional time mean density is indicated by the black contours. All data has been detrended and 5-year smoothed. Only data significant at the 99% level is shown. Significance is estimated using a moving blocks bootstrap approach (Wilks 1997) reconstructing the composites 10000 times by resampling the data with a block length estimated from the autocorrelation in each original composite.

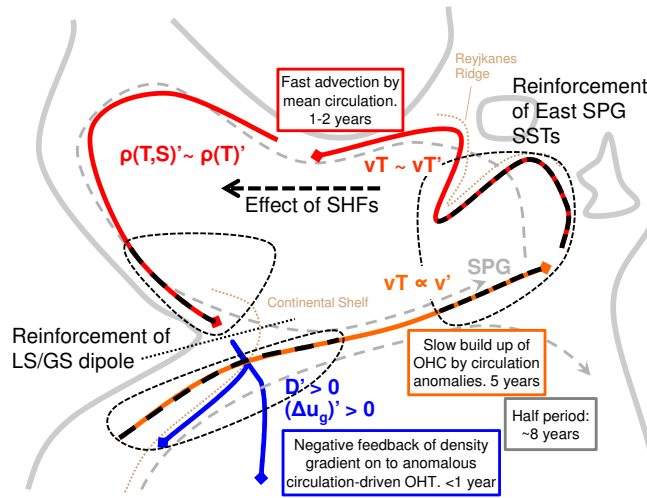


FIG. 9. A schematic of the proposed mechanism. The various processes in different regions, the timescales, and the postulated role of the atmosphere are as described in the text. Dashed grey lines denote the approximate location of NA SPG and subtropical gyres with bathymetry of particular interest marked brown. Also marked are regions where ocean heat transport correlates with circulation anomalies (orange) or is dominated by temperature anomalies (red) and where the negative feedback is suggested to occur (blue). Additionally, black dashed lines denote regions where the atmosphere is postulated to play a role in forcing or feeding back on ocean anomalies.

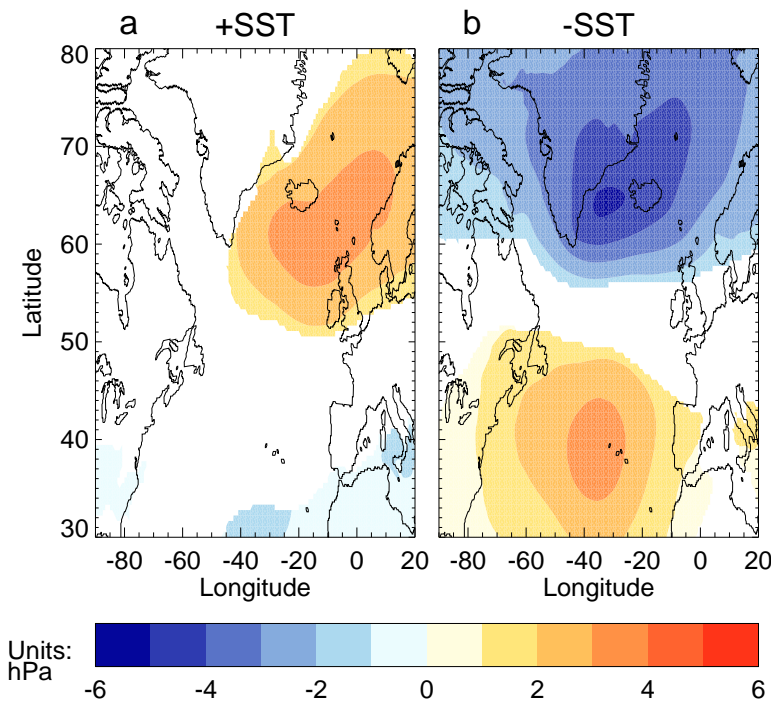


FIG. 10. Mean sea level pressure (MSLP) composites created using area-averaged eastern NA SPG SSTs (10–30°W, 50–65°) to highlight the asymmetry between positive and negative phases of the proposed mechanism. a) Composite created using the highest 10% of SST anomalies. b) As (a) but for the lowest 10% of SST anomalies. Only data significant at the 99% level is shown. Significance is estimated using a moving blocks bootstrap approach (Wilks 1997) reconstructing the composites 10000 times by resampling the data with a block length estimated from the autocorrelation in each original composite.

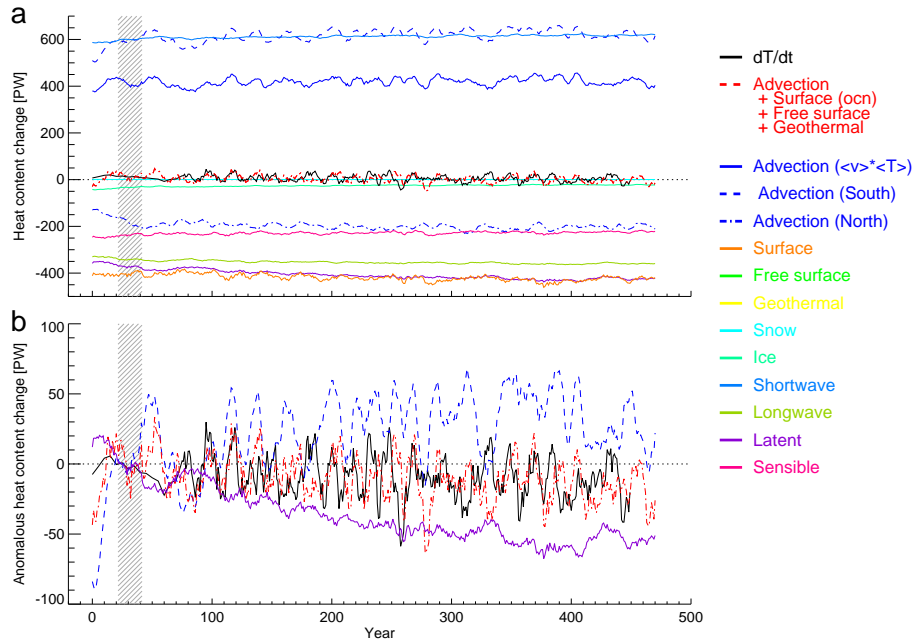


FIG. 11. The full depth heat budget of the NA SPG (53–73°N) volume plotted using 9 year running means for clarity and to highlight decadal variability. Positive is into the specified region. a) Individual components of the heat budget as denoted in the legend. b) The anomalous heat budget (referenced against years 22–42) to highlight the trends in latent heat fluxes and advective heat fluxes through the southern boundary. Note that the heat content change ($\frac{dT}{dt}$, black) and sum of heat fluxes (red) do not match prior to the year 100 as instantaneous ocean temperatures (used to calculate $\frac{dT}{dt}$) were stored with intermittent frequency during this period.